THE EFFECT OF BEDROCK GEOLOGY ON SEDIMENT YIELD IN AN ALPINE AREA OF EXTREME RAINFALL

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In this scarred country, this cold threshold land,
The mountains crouch like tigers. By the sea
Folk talk of them hid vaguely out of sight.
But here they stand in massed solidarity
To seize upon the day and night horizon

....

from The Mountains,
James K Baxter, 1961

Frontispiece: Head of the Cropp River looking south from Sentinal Peak, February 1980.

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ABSTRACT

Catchments in the Cropp River area of Central Westland are subject to very high sediment yield in an area which coincides with near maximum uplift rates and rainfall values for the Southern Alps.

In order to investigate the effect of bedrock geology on sediment yields in this area bedrock geology was mapped and a structural and metamorphic history determined, from petrographic and field evidence. Sediment yield was estimated from suspended sediment measurements and geomorphology.

The preferred interpretation of structural history is compatible with most other work done on Alpine Schist petrogenesis. It involves:

1 An early large-scale isoclinal folding (F_1) with development of an incipient fabric (S_1) parallel to bedding (S_0);
2 A pre-mid-Cretaceous regional metamorphic event (M_2) along a linear trend parallel to the present trend of the Alpine Fault and associated folding (F_2) and fabric development (S_2);
3 A probable late Cenozoic metamorphic event (M_3) affecting rocks in the western schist belt with development of a schistosity (S_3) sub-parallel to S_2 and F_3 folding;
4 Recent, possibly continuing faulting producing minor kink folds (F_4) over the width of the schist belt. The M_2 and M_3 metamorphic events occurred in a narrow zone subjected to lateral compression and probably shear heating, and in the case of M_3 this zone is related to the Alpine Fault.

Geomorphological geothermal and geological considerations suggest that uplift is occurring at present, localised on the Alpine Fault in the west and on northeast-trending faults in the east. The Cropp area may be in the zone of maximum uplift in the Southern Alps with uplift of 12 ± 2 mm per year.

Sediment yield studies: a) from a one-year suspended sediment record, and b) from reconstruction of a 12,000 year geomorphic surface and calculation of the volume of material eroded, suggests a sediment yield of 35,000 tonnes/km^2 per year, a denudation rate of 13 mm per year for the upper Cropp. Sub-catchment sediment yield is found to vary, reflecting catchment geomorphology and the rate and extent of fluvial downcutting.

Bedrock geology affects sediment yield by controlling patterns of sediment supply as a function of rock strength and therefore erodibility and controls the mechanism of sediment transport by determining the grain size of material to be transported.

INTRODUCTION

In the past 15 to 20 years there has been much research into controls on erosion in the South Island high country. However, little attention has been paid to alpine catchments west of the Main Divide. Similarly the structural history of the bedrock of the Southern Alps has been well studied in the more accessible areas in South Westland, Central Canterbury and the areas of interest to alpinists, Mt Cook and environs, but not in the Western Alps.

This study attempts to fill both these gaps by providing information on the lithologic and structural components of an area of Alpine Schists in a region which has received little study. This is combined with a semi-quantitative study of erosion processes and sediment yields in the area with emphasis on the control bedrock geology has on distribution of these processes.

The area studied was chosen because

1. it extended a programme of reconnaissance mapping in Torlesse Group greywackes through to the Alpine Fault;
2. it lies in an area of geological rapid uplift;
3. rainfall data and hydrological data collection facilities were available.

The study has four principal aims:

1. to determine the geologic history, especially recent deformation history;
2. to assess erosion processes and rates in a rapidly uplifting area of extreme rainfall;
3. to relate sediment yield values from restricted catchments to downstream agradation problems;
   and
4. to assess the validity of using suspended sediment yield as a measure of erosion in an alpine catchment.

These aims are discussed in three parts. The first in Part I of the thesis, the others in Part II.

DESCRIPTION OF THE STUDY AREA

The study area comprises the catchments of the Tuke and Cropp Rivers, Seddon Creek and Stag Creek all of which drain an area in the Southern Alps southeast of Hokitika and west of the Whitcombe River (Fig. 1). Sediment yield determinations and observations on erosion processes are from the upper Cropp River. The more extensive area was studied to determine regional geologic and geomorphic history.

CLIMATE

The climate of this area is one of its most distinctive attributes. Rainfall is extremely high with considerable variation across the area (12-8 m yr^{-1}). Precipitation appears to be relatively uniform throughout the year but poor snowfall records make this difficult to substantiate. There are more than 200 rain-days per year with 40 days of greater than 100 mm per day and 7 greater than 200 mm per day in 1979-1980.

Daily and annual temperature variation is probably small with an annual range of -5 to 25^{o}C.

More detailed rainfall records are discussed in Part II, Chapter 8.

VEGETATION

Vegetation patterns in the area are largely determined by exposure to wind and duration of snow cover. In the high alpine areas covered by snow for more than half the year vegetation is restricted to small herbaceous alpine plants. These areas are generally on the exposed ridge tops and in cirque basins.

The areas covered by winter snow, but clear in the spring and autumn are covered by various snowgrass (chionocloa sp.) communities.

On steeper slopes at similar altitudes to the snowgrass communities of podocarp cedar forest and sub-alpine scrub (oleria sp.) dominates either because of better (more youthful) soils or because avalanching clears the snow earlier in these areas. In the lower Tuke and Cropp catchments mixed conifer-hardwood forests of rata, rimu, kamahi, totara, miro, matai and pahautea occur. Throughout this forest are large areas supporting pioneer communities and even aged forest stands which represent a mosaic of revegetating slips of various ages.

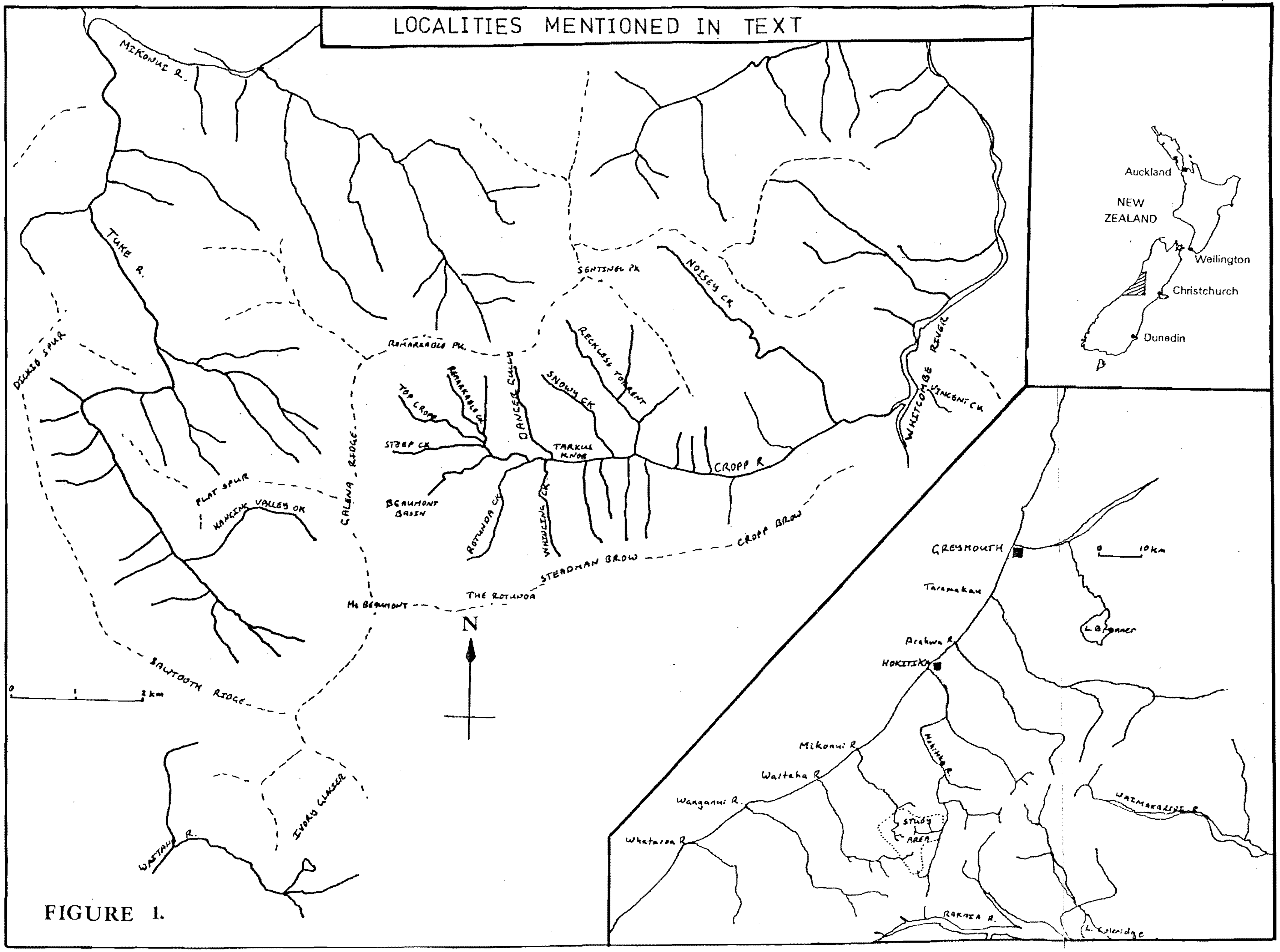

LOCALITIES MENTIONED IN TEXT
MIKONUI R.
TUKE R.
SENTINEL PK
NOISEY CK
RECKLESS TORRENT
REMARKABLE PK
DANCER GULLY
SNOWY CK
REMARKABLE CK
TOP CROPP
STEEP CK
TARKUS KNOB
CROPP R.
WHITCOMBE RIVER
VINCENT CK
BEAUMONT BASIN
ROTUNDA CK
WHIRLING CK
GALENA RIDGE
FLAT SPUR
HANGING VALLEY CK
DICKIE SPUR
CROPP BROW
STEADMAN BROW
THE ROTUNDA
Mt BEAUMONT
N
SAWTOOTH RIDGE
0
1
2 km
IVORY GLACIER
WAITAHA R.
Auckland
NEW ZEALAND
Wellington
Christchurch
Dunedin
GREYMOUTH
0
10 km
Taramakau
L. Brunner
Arahura R.
HOKITIKA
Mikonui R.
Waitaha R.
Wanganui R.
Whataroa R.
STUDY AREA
RAKAIA R.
L. Coleridge

FIGURE 1.

SOILS

A range of soils from recent through gley recents and from gleys to gley podzols and including peat and peaty soils occurs in the study area. The dominant factor controlling the type of soil found in the area is the extreme rainfall. This dominating influence is modified by the age of the ground surface and the topographic position within the landscape which affects the drainage characteristics of the site (L Basher, pers. comm. 1981).

PREVIOUS WORK

Published geological work in this area is limited. The most complete regional survey covering this area of the Alps is the Mikonui Bulletin of Morgan (1908).

Morgan recognised the Alpine Fault as a major overthrust of considerable importance in the uplift history of the axial ranges, and that the alpine folding in the western alps (schists) post-dated folding in the eastern graywackes.

The NZGS 1:250,000 geological map of the area (Warren 1967), provides little data and less detail than Morgan's maps.

Work is currently being undertaken on the geochemistry of the Pounamu ultramafics in and to the north of the thesis area by Dr A Reay and Dr A Cooper of Otago University. Koons (1978) studied the metasomatism of the Pounamus in an area of similar rocks 10 km to the northeast.

Hawkes et al. (1980) and Bromell et al. (1980) produced maps of the Torlesse Group rocks in the nearby Rakaia-Whitcombe Rivers area.

Relevant studies on structure and metamorphic history of Haast schists include: Grindley et al. (1958) on the structure of the Trent River region; Grindley (1963) on metamorphism and structure of the Haast River region also Cooper (1976) on the Haast area; Ward and Spörli (1979) on structural history of the Mt Cook region; Andrews et al. (1974) in the Lord Range and Findlay (1980) on the Alpine Schists in the Copeland area of South Westland. All of these studies are either of low grade Torlesse Group rocks or of polymetamorphic rocks in South Westland.

Other studies relating erosion to geology include J Adams (1978) thesis on alpine erosion in the South Island which gives a very general account of uplift and erosion; O'Loghlin (1969) related streambed morphology and erosion history in a small Torlesse catchment; Blair (1972) related erosion patterns to lithology and structure in a catchment in the Torlesse Range.

McSaveney (1979) quotes studies of sediment yield in schist catchments in Otago and claims that sediment yield of alpine catchments is related to rainfall and largely independent of geology. Also that by far the majority of the load in alpine streams is carried as suspended load.

Hayward (1980) records detailed sediment yield studies for a small Eastern Alpine catchment where most of the sediment is transported as bedload.

PART I

THE GEOLOGY OF THE CROPP RIVER AREA

CHAPTER ONE: SYNOPSIS

The study area lies within the main schist belt of the axial ranges of the Southern Alps. Metamorphic grades range from chlorite zone (IIA) to oligoclase zone. Rocks of the area are classified within the Haast schists (Warren 1967; Cooper and Reay 1977). However, the metamorphic history and resulting textures and mineralogy are sufficiently different to the Otago schists that they are better described by the informal term Alpine schists.

Most of the schists of the area are quartzofeldspathic texturally similar to nearby Torlesse Group rocks. In addition there is a range of basic and ultrabasic metavolcanics, metacherts and marble in an apparently simple stratigraphic succession.

The ultramafic rocks belong to the Pounamu Formation (Morgan 1908). They consist of sill-like pods of serpentine-dunite and serpentinite structurally conformable with the schists and surrounded by a metasomatic aureole. Structural evidence suggests that tectonic emplacement of the ultramafics preceded the metamorphic maximum.

Structural observations, microstructural evidence, K/Ar dating and textural mapping in the Cropp-Tuke area result in the adoption of a structural scheme involving three folding and three associated metamorphic events.

The earliest structures are large-scale probably isoclinal folds (F_1) in bedding with mimetic recrystalisation (M_1) enhancing bedding, locally producing a S_1 fabric.

The second phase of folding (F_2) refolded S_0 and S_1 about steeply plunging fold axes into near vertical isoclines and was both accompanied and post-dated by metamorphism (M_2) to oligoclase grade. This event occurred before 70 Ma and was probably associated with a linear shear zone similar to the present Alpine Fault system.

A third phase of folding (F_3) is developed only in the western schists. It is coaxial with F_2 and associated with high grade metamorphism in the west and retrogressive metamorphism in the central schist belt. This metamorphic event occurred pre-5Ma and is considered to be a late Tertiary event associated with uplift of the present Southern Alps.

Three main fault systems are present in the area. North and ENE-trending faults are probably conjugate shears, recently active in response to present tectonic compression. A conspicuous NNE-trending fault zone following the eastern boundary of the high grade schists probably pre-dates the recent fault systems.

Because of the complex metamorphic history and the related complexity in geothermal gradients interpretation of uplift rates from K/Ar dating is difficult but total uplifts up to 12 km are implied.

The area is in a state of geomorphic disequilibrium with rivers downcutting actively into a landscape which shows evidence of two glacial planation events. Geomorphic evidence from outside the study area suggests that the Hokitika River region is the region of maximum uplift in the Southern Alps with an uplift rate of 12 ± 2 mm yr^{-1}.

CHAPTER TWO: BEDROCK LITHOLOGIES

In general the rocks of the Cropp area are high grade metamorphics of greenschist to amphibolite facies grade.

The distribution of lithologies across the area suggests that they have retained a large degree of stratigraphic continuity. There is a predictable lithologic sequence normal to the schistosity interpreted as representing original stratigraphy. The most convincing individual bed in the sequence is a marble band traceable for 8 km+ through the Cropp and upper Tuke catchments.

Six distinctive lithologic groups are mapped (see Map 1)., the following is a description of their occurrence and petrology.

Metapsammite

Metapsammites are the coarser grained (0.1-1 mm) macroscopically usually, microscopically always schistose quartzofeldspathic rocks which form the major component of the bedrock through most of the mapped area.

The major constituent of these rocks is quartz with feldspar as a fine-grained allotriomorphic groundmass. The major accessory minerals are chlorite, biotite, epidote, and garnet developed in various proportion depending on metamorphic grade. Minor accessories include stilpnomelane in the lower grades, apatite, sphene and clinozoisite.

Although mean grain size in this lithotype is silt or fine sand sizes, relict detrital grains in some sections show an original medium-fine sand range (Fig. 2). These rocks are probably the correlatives of the sandy textured Torlesse Group rocks to the east. In the amphibolite facies rocks on the western part of the area grain size increases with increasing mineral grade and relict textures are lost to metamorphic textures.

Metapelites

Metapelites are the fine-grained equivalents of metapsammites. Grain sizes, where discernible are very fine (0.05 mm$\pm$) and colours are darker than for psammites. There is a complete gradation between these rock types.

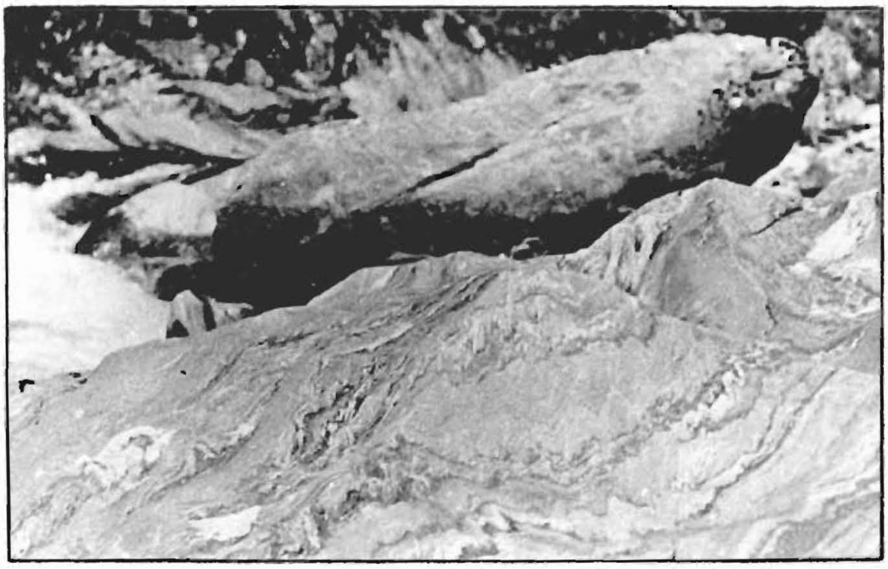
As far as can be determined optically metapelites are compositionally similar to psammites although graphite and pyrite are more often found in the finer rocks.

Figure 2: Psammitic schist. (a) Lower Cropp River, garnet zone. (b) Mid-Tuke River, garnet zone.

Figure 2c: Photomicrograph showing texture of psammitic schist. Note the large quartz and feldspar grains preserving original grain size in a groundmass of fine-grained "annealed" quartz with biotite porphyroblasts.
(U of C № 9554)

a

b



Metacherts

Metacherts are generally fine textured, milky white rocks. They often occur in discrete "beds" interbedded with metapelites. Compositionally metacherts are dominantly quartz with 10% to 30% albite. Superimposed on a granoblastic matrix of quartz and feldspar, often with a strong optical preferred orientation is a scattering of micas with a strong dimensional preferred orientation. The micas are usually in two populations; fine plates of muscovite and/or chlorite defining schistosity with generally larger biotite porphyroblasts. In some cases garnets may be present as late stage porphyroblasts.

The metacherts are commonly iron-stained on the surface and occasionally are associated with sulphides. In one case large (2 cm) euhedral pyrites are present (Fig. 3) in another a massive pyrrohtite-pyrite-marcasite association occurs in 5 cm by 1 m stringers. In both cases the sulphide-bearing layers are concordant with the surrounding strata. Cummingtonite-hornblende-garnet-bearing metacherts are found where cherts and metavolcanics are in close association.

The composition and stratigraphic associations of these metacherts suggests that they are related to the manganese, iron-rich red and green cherts of the Torlesse Group. Presumably the metal ions have accumulated in mineral-rich zones during metamorphism cleaning the colour out of the chert.

Marble

Marble is present in minor proportions associated with metavolcanics but the main occurrence is a discrete bed of coarse-grained (0.1 to 4 m) yellow-white marble. This rock type is a schistose marble with granoblastic calcite interspersed with muscovite and phlogopite and occasionally with quartz lamelli.

Marble occurs associated with metapelites and metavolcanics and the same bed can be traced for 8 kilometres. As with the metacherts this bed is not continuous but is tectonically stacked where it crosses fold hinges and thinned or even lost on fold limbs (Fig. 4).

Limestones are known from the Torlesse terrain but this deposit represents a considerable volume, more than would be expected considering the paucity of known limestones in the Main Divide region.

Figure 3: Large pyrite crystals in chert with disseminated pyrite throughout the groundmass. Sulphides appear to be restricted to discrete beds usually in chert dominated sequences.

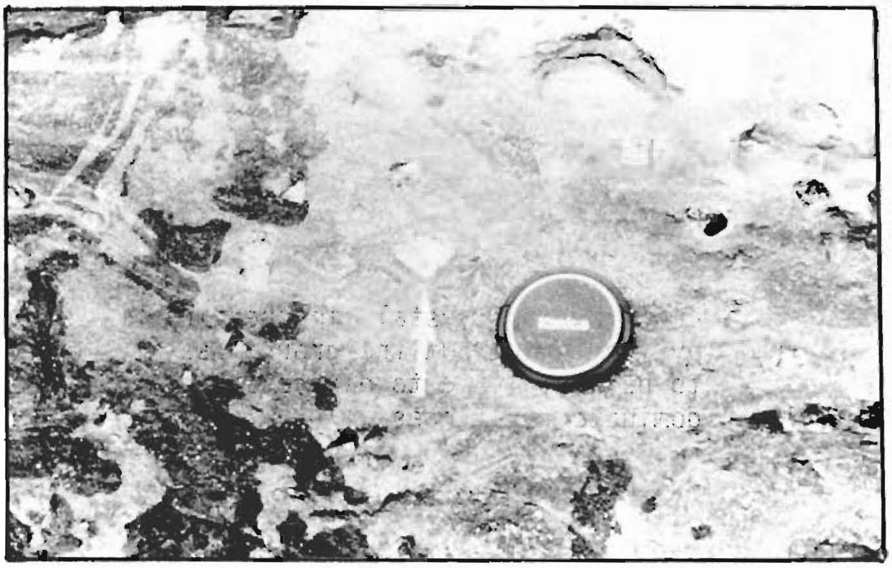


Figure 4: Marble. (a) Marble bed on Galena Ridge, up to 5 m thick where tectonically stacked across mesoscopic fold hinges. (b) steeply dipping marble, Steep Creek. (c) Rootless folds with thin marble in hinges.

Figure 5: Thin greenschist bed, Warning Creek.

a

b

c

Metavolcanics

Two classes of metavolcanic rock are represented in the area - greenschists and ultramafics.

Greenschists occur concordant with other lithologies in the quartzofeldspathic schist (Fig. 5). They are texturally similar to the metapelites but have a greyish green colouration due to the abundance of chlorite. Greenschists have not been mapped into the amphibolite facies, partly because they loose their distinctive colouration as chlorite converts to amphibole and partly because of their limited distribution.

Petrographically the greenschists are dominated by fibrous chlorite forming the schistosity planes and dividing off augen of quartz, feldspar and calcite segregation lamellae. Accessory minerals include epidote, biotite, garnet, hornblende and clinozoisite commonly with pyrite.

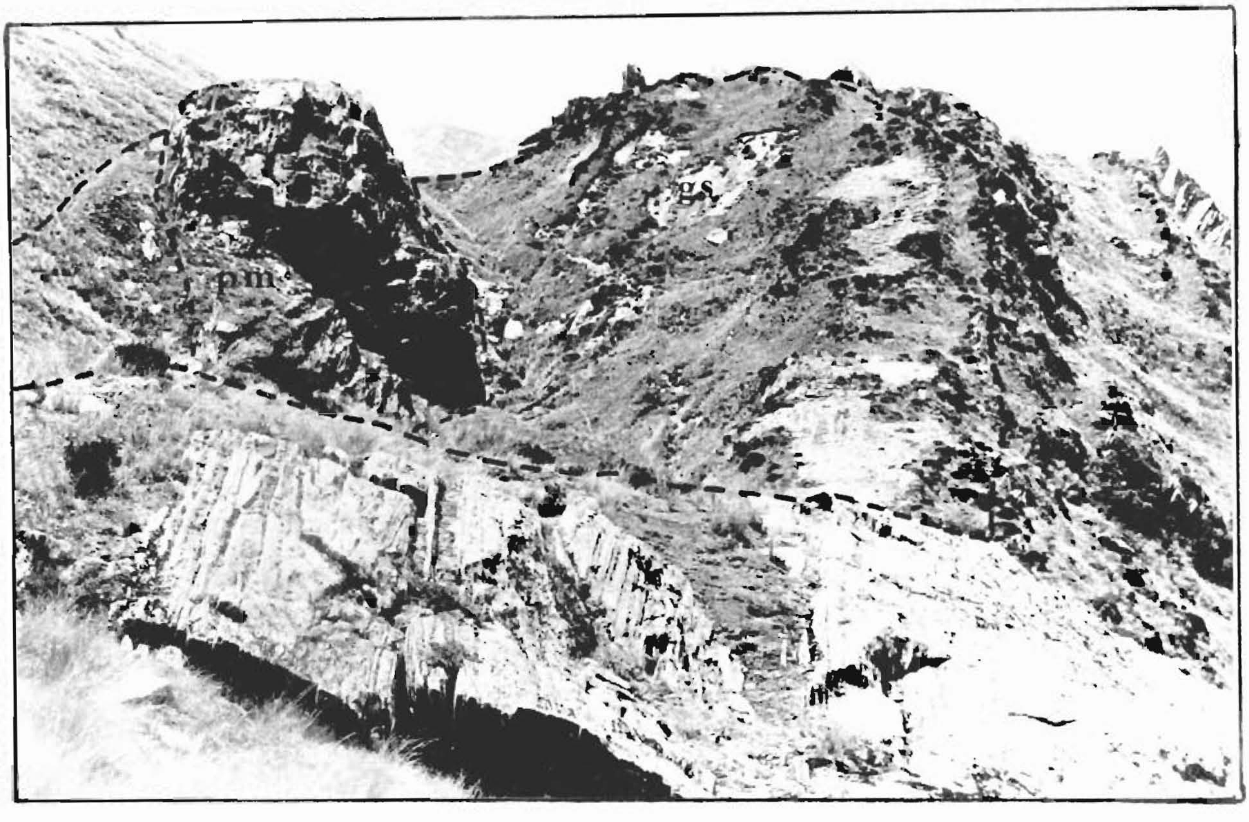
Greenschists are volumetrically important in the upper biotite zone of the study area comprising 20% to 40% of rocks exposed. It is commonly assumed that greenschists of the Haast schists are metamorphosed pillow lavas and volcaniclastic sediments. The presence of volumetrically significant volcanogenic material in the Cropp succession contrasts with typical Torlesse low rank assemblages which have only rare volcanogenic beds.

The area is also distinguished from a typical Torlesse succession by the presence of three belts of ultramafic rocks in the northeast of the area (Fig. 6) belonging to the Pounamu Formation. These rocks are described by Morgan (1908), Cooper and Reay (1977) and Cooper and Reay (in prep.). The metasomatic history of similar rocks 15 km to the north is dealt with in Koons (1978).

Basically there are two well exposed elongate pods of serpentine-dunite and serpentinite surrounded by envelopes of talc-magnesite and greenschist. Brecciated amphibolite is locally present on the margins. The serpentinite core is cut by anastomosing zones of high strain parallel to regional schistosity. Large euhedral magnetites and pyrites are locally found in the margins of the ultramafics.

There is no unmetamorphosed analogue to the Pounamu ultramafics immediately available. Cooper and Reay (1977) consider that they may represent a slice of oceanic crust tectonically emplaced before the major deformation. This point is discussed further in Chapter Four.

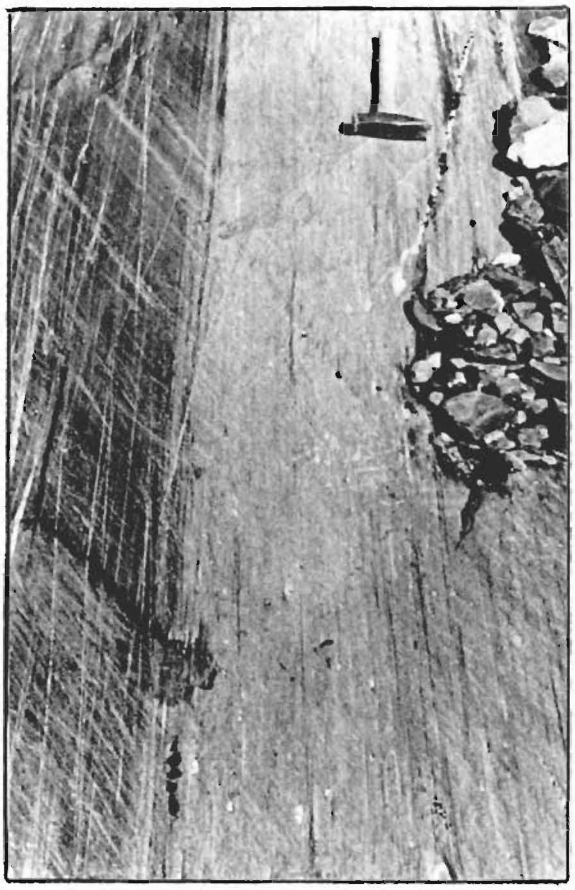
Figure 6: Pounamu Ultramafics, Noisey Creek. pm = serpentinite and serpentine-dunite core enveloped by a sheath of metasomatised greenschist (gs), surrounded by quartzofeldspathic schists.


pm
gs

Numerous gradations between these general groups are present in the Cropp area. Pelites and psammites are distinguished on texture of which all variables exist including rare conglomerates (Fig. 7). Metachert is distinguished by its low content of mafic minerals. Greenschist is mapped on the basis of colour, and tectonic mixing makes boundary placement uncertain.

With all features considered the rocks of this area represent a metamorphosed stack of Torlesse type sandstones, mudstones and cherts interspersed with basic volcanics and/or volcaniclastics. Within this pile was a limestone of significant size. The whole mass was probably not too far distant from the primitive basin floor allowing inclusion of ultramafic pods during deformation.

Figure 7: Conglomerate bed, Ivory Glacier. Note pebbles highly elongate parallel to schistosity. Schistosity is parallel to bedding, lineations oblique to bedding are glacial striations.



CHAPTER THREE: METAMORPHIC HISTORY

3.1 METAMORPHISM

The rocks of the study area do not show a simple progression of textural and mineralogical zones as is found in Otago (Bishop 1972) or Haast (Cooper 1974) or in the central Southern Alps (Warren 1967; Gair 1967). Metamorphism appears to have involved three temperature maxima and the resulting textural zonation is complicated by the overprinting of these events which vary in extent and degree over the area.

Various models for the metamorphic history of the area are discussed in the light of regional geological considerations in section 3.6 and a model involving three events is proposed. These events are: a pre-Cretaceous localised metamorphism associated with large-scale folding; a widespread high grade Cretaceous event, and a late Tertiary event involving high strain and high geothermal gradient.

3.2 METAMORPHIC FACIES

Metamorphic facies are mapped using isograds defined by the first appearance of key minerals in rocks of quartzofeldspathic composition. Metastable assemblages are common in the area and this makes mapping of isograds difficult especially in areas which appear to have undergone retrogressive metamorphism.

Because metamorphism may not have gone to completion in all areas zonal boundaries are either indistinct, gradational, or complex. As a result of this combined with often widely spaced sample points, the exact shape of the isograds is not well determined. In general isograds are positioned within 100 to 500 m and dips are generally steep (see Map 2).

Prehnite-Pumpellyite facies

The low grade equivalents of rocks in the study area are the quartzofeldspathic Torlesse Group rocks of prehnite-pumpellyite facies to the east and south of the area. These rocks are close to their original clastic composition and texture with metamorphic minerals restricted to matrix and veins.

Clastic composition is assumed to be similar to the quartzofeldspathic greywackes described by Andrews et al. (1973) (Table 1).

Greenschist Facies: Chlorite Zone

The appearance of metamorphic chlorite marks the onset of greenschist facies metamorphism. The chlorite isograd is not mapped but is thought to run northeast through the Wilkinson Valley. Local areas of chlorite zone are present to the east associated with major faults and fold hinges in low grade rocks.

The chlorite zone has been used as a mapping unit in preference to the pumpellyite-actinolite and pumpellyite-clinozoisite zones. This is because of the small number of samples collected from the low grade rocks and the overwhelming dominance of late stage oriented chlorite in the samples collected.

Bedding features and relict detrital grains are present throughout the chlorite zone but with a fine mortar texture of recrystallised very fine-grained quartz and a variably developed fabric defined by chlorite with white mica, stilpnomelane and clinozoisite.

In areas of low grade rocks to the east and south of the Cropp area a mimetic recrystallisation parallel to bedding has occurred associated with early folds. In the chlorite zone of the study area this recrystallisation is the dominant textural feature.

Biotite Zone: There are two separate areas of biotite zone mapped. These are mineralogically similar but texturally quite distinct.

In the east biotite appears growing with chlorite and within a short distance becomes the dominant mica. Within the eastern biotite zone two generations of biotite occur, a synkinematic growth along schistosity and a post-kinematic porphyroblast phase.

Albite is the only feldspar in this zone, often occurring as relict detrital grains and often twinned. Bedding is well preserved and graded beds are common.

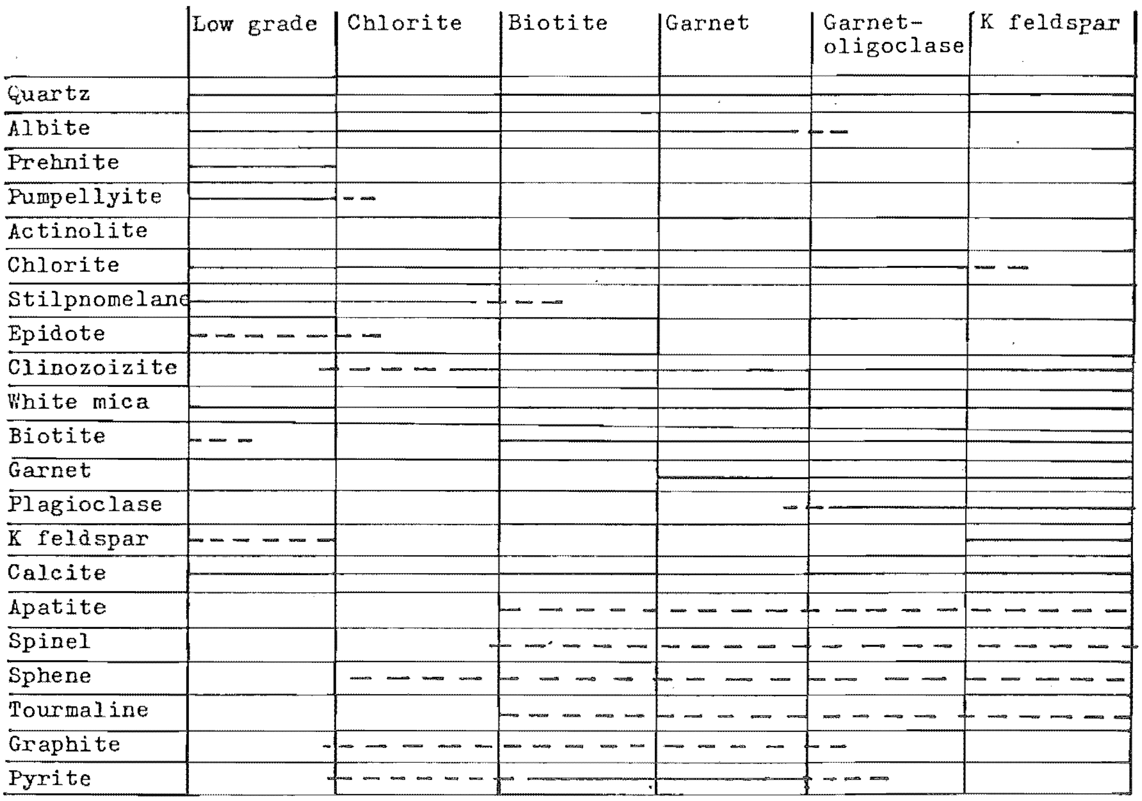


TABLE 1:Mineralogy of quartzo-feldspathic (psammitic) schists of the Cropp River area.After Findlay (1980).

The western biotite zone on the other hand has no detrital grains and three generations of biotite. As well as the early synkinematic and post-kinematic biotite phases there is a post-porphyroblast phase of schistosity producing biotite (and chlorite). The feldspar is albite and accessory minerals include chlorite, clinozoisite, calcite, white mica and opaques.

Bedding is completely transposed in this zone and early structures are deformed.

Garnet Zone: The garnet isograd is mapped on the first appearance of almandine in quartzofeldspathic schist. Three areas of garnet zone are mapped, each separating areas of biotite zone from higher grade oligoclase zone rocks. In the east garnet zone rocks flank a narrow zone of oligoclase zone and in the west they occur in a wide band showing progressive westward increase in grade.

The eastern garnet zones are narrow (~1 km±) and the rocks are texturally indistinguishable from their equivalent biotite zone rocks. They show graded bedding and have detrital grains of quartz and albite (which may be twinned). The zone is defined by small porphyroblastic almandines scattered through the rocks.

In the western zone garnets increase in size with increasing grade. There has been a complete metamorphic recrystallisation in these rocks with no sign of remnant detrital texture.

Amphibolite Facies: Oligoclase Zone

The presence of metamorphic oligoclase indicates amphibolite facies conditions. Again two separate belts of oligoclase-bearing rocks occur. In the east oligoclase-bearing rocks differ from garnet zone rocks only in that oligoclase replaced albite. Albite and oligoclase coexist in several samples and in one case oligoclase overgrows a detrital albite grain. Garnets never develop significant size in this zone, the largest observed being less than 3 mm.

In the western belt garnets become progressively larger (up to 40 mm) and the degree of textural modification, development of mineral segregation laminae, increases toward the west. Albite and oligoclase are not known to coexist in this zone.

<u>High Grade Zone</u>: In the far west (Dickie Spur) is a zone or series of zones of rocks with high grade mineral assemblages. These rocks have no garnets, a high Na-oligoclase and K-feldspar (with quartz, biotite and muscovite).

The K-feldspar is indisputably metamorphic occurring with a web-like texture between quartz and plagioclase grains (Fig. 8).

Energy dispersive microprobe observations (on the JEOL 5MX, Department of Mechanical Engineering, University of Canterbury) showed the alkali feldspars to have very high K and very low or no Na. Crystal habit suggests that these feldspars are not adularia which usually is of hydrothermal origin, so are probably orthoclase and possibly sanidine.

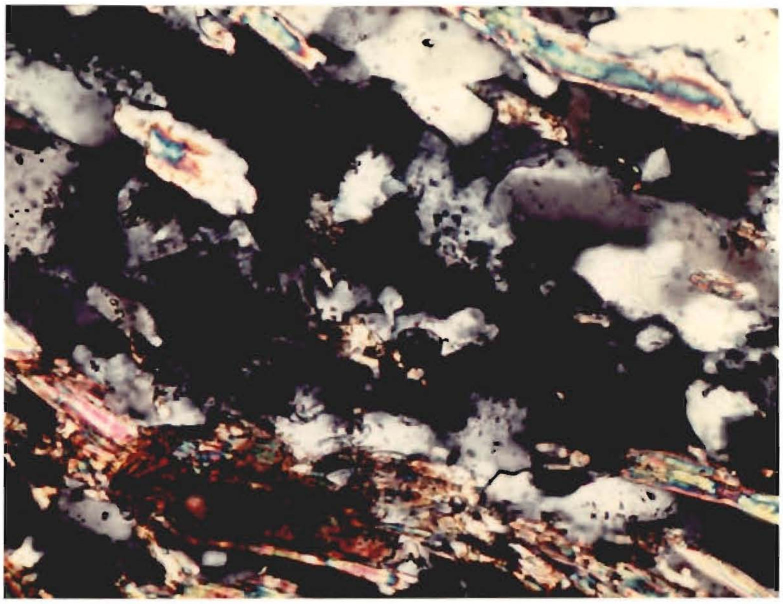
3.3 TEXTURAL ZONES

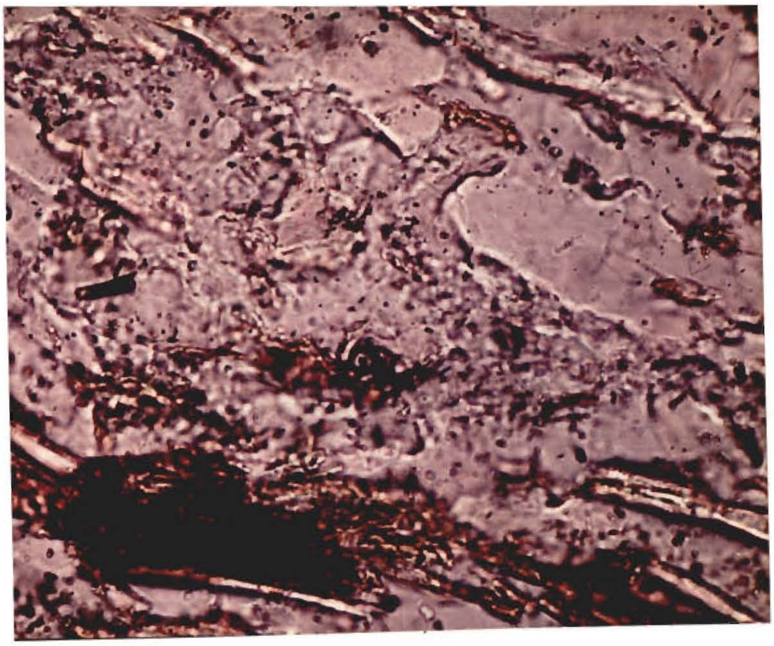
The scheme for mapping the degree of development of metamorphic textures developed for the Otago schists (Bishop 1972) and applied in the Haast-Mt Cook area (Cooper 1974; Findlay 1980) produces two apparent anomalies which applied to the study area.

Firstly textural zonation is complicated by the development of isolated narrow zones of high textural grade. In these zones a high degree of mineral mobilisation and differentiation has occurred producing mineral segregation parallel to bedding and enhancing bedding features and raising the textural grade from a regional II (A or B) value to IV. These narrow zones of high textural grade are similar to zones in the West Mathias River where textural zone IV rocks are developed in the core of a large early fold (personal observation) and in the Garden of Eden area, associated with early inclines (J Bradshaw, pers. comm. 1981). The rare areas of high textural grade are superimposed on an overall pattern of increasing grade toward the Alpine Fault.

Secondly there is a major variation in the relationship of textural and mineralogical grade between the Cropp and other mapped areas in the Haast schist. In this area IIA and IIB textural zones continue from chlorite through to oligoclase grade rocks with prominent mineral segregation laminae (IIIA, IIIB) not developed regionally until the western garnet zone. Within a kilometre of the Alpine Fault further grain size reduction presumably associated with cataclasis reduces the grade to IIb but this is not mapped (see Map 2).

Figure 8: Photomicrographs of K-feldspar-bearing schists from Dickie Spur. (UdC No 9552)





Disparity between this and other areas must be due to differences in the nature of the metamorphic events. That is, to temporal and spatial variations in the factors which cause alignment of platey minerals (zones IIA, IIB) and development of segregation laminae (IIIA, IIIB).

Factors which could constrain mineral growth in a regional metamorphic system with a range of P-T conditions and high total strains in quartzofeldspathic rocks are difficult to conceive. One possibility is that there was insufficient time available for textures to develop. This is unlikely as high textural grades are developed in related rocks of low mineral grade 10 km to the south in the Lord Range (Andrews et al. 1972).

The other possibility is that a suitable aqueous medium was not present in the rocks during the metamorphic event. Prograde metamorphic reactions produce water which is then available to catalyse mineral growth and grain boundary reactions within the rock. If fluids were removed from a rock undergoing metamorphism it is possible that crystal growth would be reduced. A mechanism by which fluids can be removed is suggested by Fyfe et al. (1979). "Rocks undergoing metamorphic reaction are frequently schistose or exhibit cleavage . . . If such planes of anisotropy (as cleavage) are steeply inclined they will form a direction of relatively easy migration and so greatly facilitate upward movement of fluids."

If the schists of the study area had near vertical schistosity and bedding during metamorphism fluids could have migrated upwards causing the zone of potential mineral growth to be depleted in fluids necessary for crystal growth. That is not to say the metamorphic system was dry but that PH_2O was low and development of mineral segregation laminae was reduced as a result.

This hypothesis relies on the assumption:

1 that metamorphic fluids can be removed from the reacting system by migration along grain boundaries with only minor veining;
2 that schistosity and bedding were nearly vertical at the time of the main metamorphic event and thus provided no traps to fluid migration.

The hypothesis can be tested by looking at mapped areas where schistosity and bedding vary from steep to gently inclined synmetamorphic orientations. It would be expected that mineral segregation would be well developed in flat lying areas where bedding and/or schistosity would form traps inhibiting fluid migration and less well developed in steeply dipping areas.

In the Lord Range flat-lying bedding is intersected by a steeply dipping schistosity with higher textural grades than are observed in the study area. This is possibly a case of bedding acting as a barrier to fluid migration resulting in higher textural grades. In the Otago schists both bedding and schistosity are sub-horizontal and textural grades are high. Findlay (1980, p. 125) describes large steep folds in Alpine schists in the Fox Glacier region where the steeply dipping limbs in high grade rocks have an axial plane schistosity defined by phyllosilicate orientation possibly without any quartz and feldspar growth. These limbs form areas of textural grade IIB where the hinge regions of the same folds are IIIB-IV. This could well be an example of near parallel steeply dipping foliations allowing fluid migration where cross cutting foliations produce fluid traps resulting in higher textural grades.

The textural scheme of Bishop (1972) is descriptive but in this area the characteristics described by the scheme have genetic implications. Mapping of textural zones produces patterns of isotects which are directly related to regional structural history. Because of this I have adopted a system of interpretive subscripts attributing the maximum textural grade present at a particular site to a particular metamorphic event. Thus a rock which underwent significant mineral segregation development during M_1 (IIIB) and no subsequent modification to a higher textural level would be mapped at $IIIB_1$. In contrast a rock which was not significantly modified during M_1 but which developed mineral segregation lamination during M_3 would be a $IIIB_3$.

Without such a modification mapping of textural zones in this area would either grossly oversimplify the true picture if done on a large scale or present a highly complex and misleading view if mapped on a small scale. A description of common textures is included as Table 2.

Because the complex relation of textural and mineral zones was only detected after considerable mapping was completed the boundaries mapped (Map 3) are only accurate to less than 500 m. However, the changes they represent are real.

TABLE 2.
Textural zones for psammitic schists of the Cropp River area as mapped in Map3. As textural characteristics are determined by metamorphic history, they are described according to the metamorphic event which dominated in determining texture.

	M_1		M_2		M_3
IV_1	Well developed mineral segragation laminae (greater than 10mm long and 2mm thick) developed without preservation of a microfabric. Laminae are probably a result of mimetic recrystalisation parallel to bedding.			IV_3	Quartz-plagioclase mineral segragation laminae (parallel to S_3) greater than 2mm thick. Complete metamorphic recrystalisation
$IIIB_1$	Mineral segragation laminae greater than 10mm long but less than 2mm thick laminae not associated with obvious schistosity and generally parallel to bedding			$IIIB_3$	Mineral segregation laminae parallel to S_3 Greater than 10mm long but less than 2mm thick Complete recrystalisation
					Mineral segragation laminae less than 10mm long. Complete metamorphic recrystalisation.
		IIB_2	Schistose sandstone splits into parallel sided slabs. Bedding often graded, detrital grains preserved.	IIB_3	Schistose sandstone, bedding features obscured. transposed whisps of rootless folds common. detrital grains rare.
		IIA_2	Weakly schistose fissile sandstone		
		I	Non-schistose sandstone		

3.4 METAMORPHIC FEATURES IN QUARTZOFELDSPATHIC SCHISTS

The lowest grade rocks mapped lie in the east of the area, in faulted contact with low grade Torlesse Group metagreywackes. They are metamorphosed to chlorite grade with textural modification limited to grain size reduction in the quartzofeldspathic matrix and, in local areas, mimetic recrystalisation parallel to bedding. Original bedding is preserved and textures are largely detrital.

Chlorite gradually gives way to biotite, usually with two generations of biotite, a fine schistosity producing phase and a later porphyroblastic phase. Chlorite occasionally grows as a late phase in kinks in biotite porphyroblasts and associated with quartz veins or with an ill-defined fabric as fine grains in the matrix. Original detrital grains are still recognisable with twinned albite and quartz showing biotite beards in pressure shadows. Graded bedding is commonly preserved (Fig. 9).

The first appearance of garnets is followed closely by the first appearance of oligoclase instead of albite. Garnets occur in rocks texturally very similar to the biotite grade rocks but for a scattering of euhedral garnets. Albite and oligoclase commonly coexist and albite may be twinned. Regressive chlorite is sometimes found forming a late stage halo around garnets and replacing biotite.

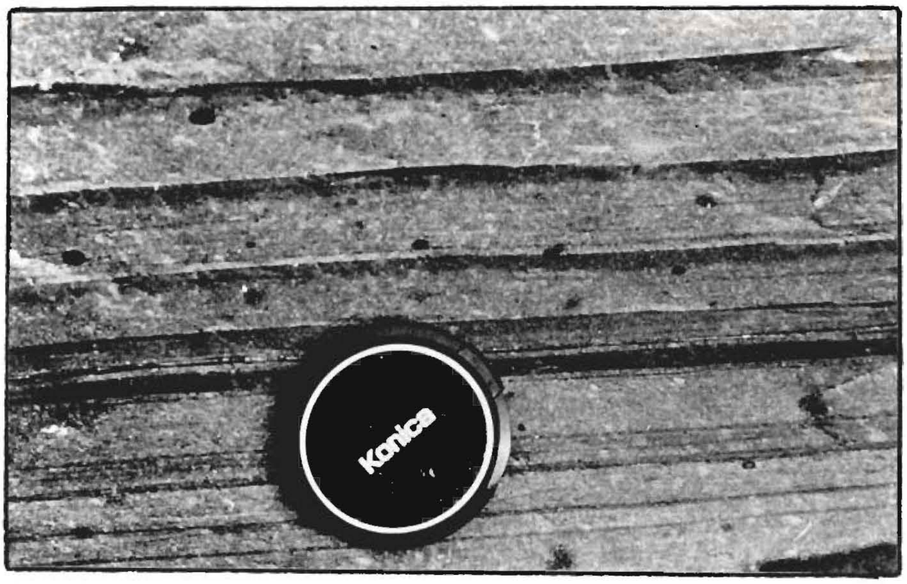
Further west there is a transition into rocks of biotite grade. These rocks are texturally distinct from the rocks to the east. The porphyroblastic biotite phase is present, post-dating the initial schistosity but cut by a later biotite-chlorite fabric. The porphyroblastic biotite has beards of later biotite in pressure shadows (Fig. 10).

In the field these rocks can be identified by their black colour and fine grain size. They include stringers of coarse-grained quartz-rich "bedding", highly folded and enclosed by black, highly fissile, often shattered schist (Fig. 11). These biotite zone rocks persist for 1 to 2 km across strike with little change except that the second schistosity (S_3 formed on late biotites) becomes dominant.

A gradation into garnet then garnet-oligoclase grade rocks occurs to the west. In this zone the complete metamorphic recrystallisation of the previous zone develops into mineral segregation laminae (textural zones III and IV) parallel to S_3.

Figure 9: (a) Graded bedding in schist of oligoclase grade. (Dark spots are raindrops, an almost ubiquitous feature of outcrops in the Cropp area.)
(b) Thin section showing partial preservation of detrital grains. (U.S.C. No 9553)

a



b

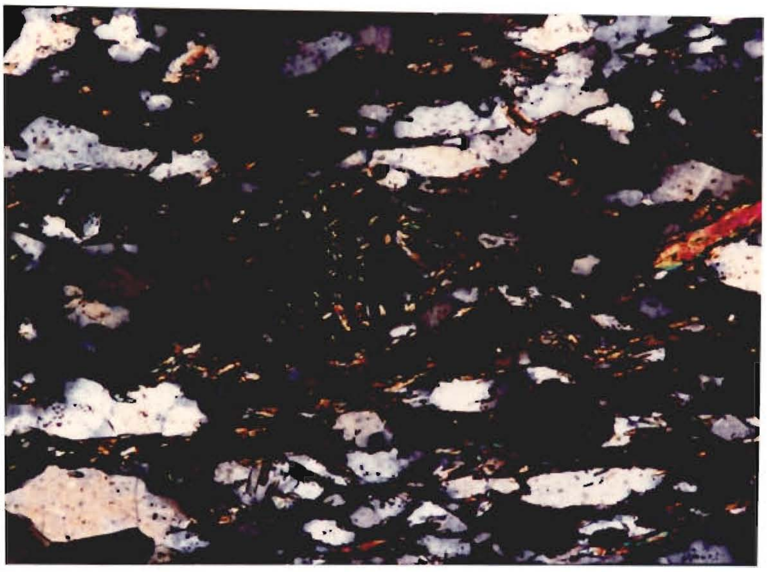
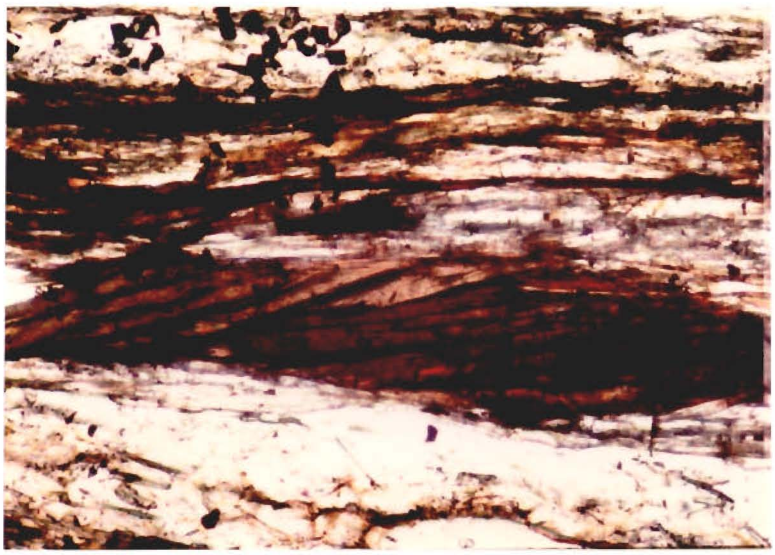


Figure 10: Photomicrograph of garnet grade pelitic schist showing porphyroblastic biotites surrounded by late stage schistosity producing biotite and chlorite. Some late stage biotite has grown in the pressure shadow of the porphyroblast.

(U of C Nos 9555 + 9559



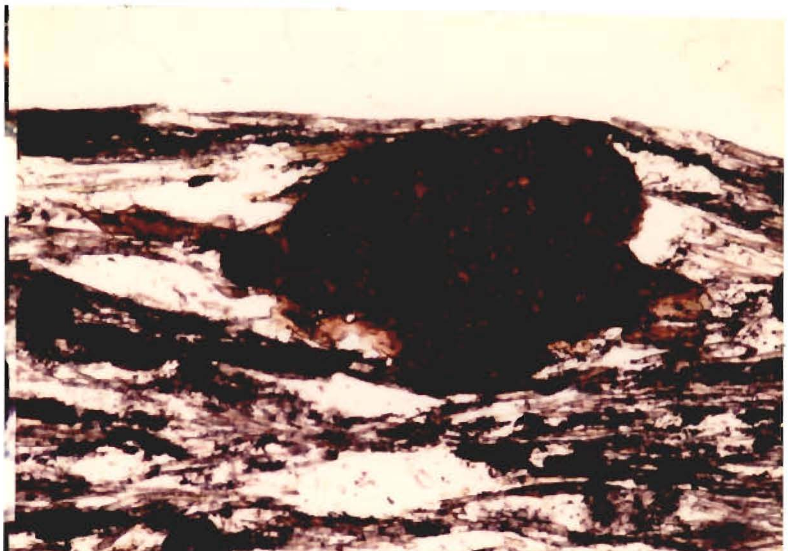


Figure 11: Highly sheared pelitic schist in the zone where S_3 begins to dominate over S_2. Upper Cropp River.

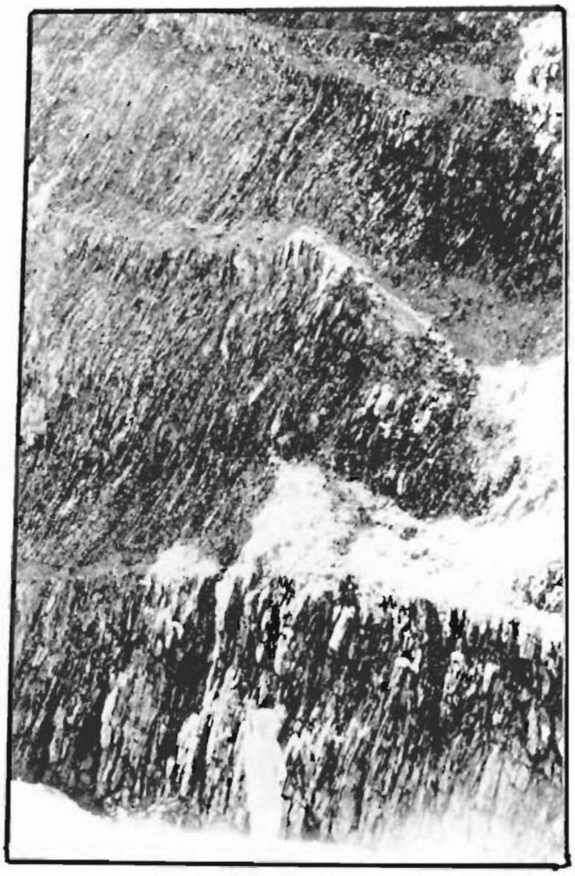
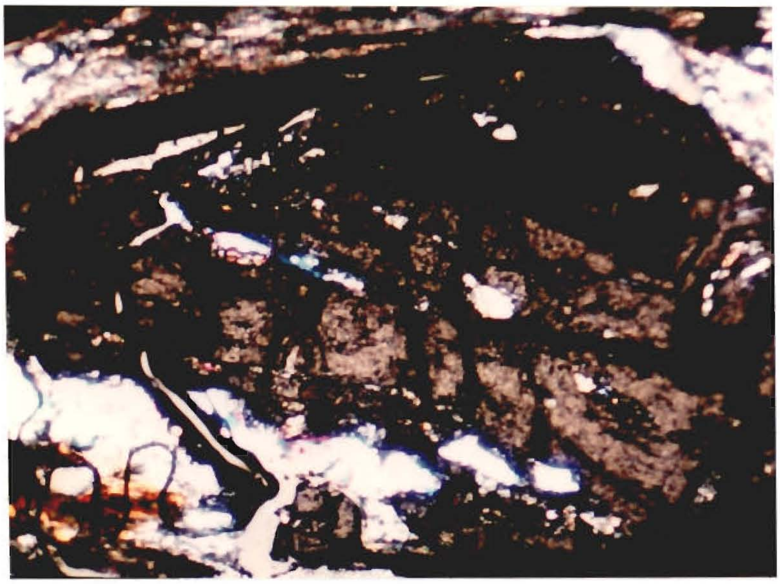


Figure 12: Garnet prophyroblast from oligoclase zone. Flat Spur in the Tuke River. Prominent schistosity is S_3, possible F_2 fold is preserved as an inclusion trail in the garnet (thin section No 116).



Post and syntectonic garnets increase rapidly in size toward the west. Garnet growth appears to accompany fabric development but with little rotation evident. In one slide an early fold hinge is enclosed by a garnet porphyoblast (Fig. 12).

Local K-feldspar-bearing rocks in the far west may represent a high temperature mineral assemblage developed during the last metamorphic climax.

The overall pattern involves three belts of metamorphic zones, an eastern area with mineralogically high grade but texturally poorly developed schists and a western zone of mineralogically and texturally high grade schists. Dividing these zones is an area of mineralogically and texturally low grade rocks.

3.5 K/Ar DATING

The samples dated were from widely scattered points in the study area. The aim of the dating study was to develop an idea of cooling age distribution across the area and its relation to structural and metamorphic features. Techniques used are presented in Appendix I and the results are included in Table 3.

Sample DS1 is from Dickie Spur about 1.8 km from the Alpine Fault within the high grade facies rocks of the western metamorphic zone. It is a quartz, oligoclase, K-feldspar, biotite, muscovite, epidote schist with one well developed schistosity formed by large biotites.

The age derived for DS1 is 2.1 to 2.7 Ma (biotite). The feldspar age of 11.1 to 12.7 Ma is thought to reflect the presence of excess argon or, more probably, a higher argon retention temperature for feldspar than for biotite. It may also be affected by the presence of K-feldspar in the sample.

DS2, from Tuke Hut, 5.6 km from the Alpine Fault in the western oligoclase zone is a quartz, oligoclase, biotite (clinozoisite, calcite, muscovite, chlorite) schist with biotite as porphyroblasts and as fine laths forming schistosity.

As for DS1 the plagioclase age is not used for DS2 because of the suscept-ibility of plagioclase to having excess argon and the possibility of high argon retention temperatures increasing the apparent age. The biotite and

TABLE 3:K Ar dates

INS No.		SAMPLE No	ROCK TYPE & LOCATION	Grid Ref.	K(av)	40K/36Ar X1000	$^{40}Ar/^{36}Ar$	40Ar(rad) nl/gm	%tot	AGE Ma
7560	Bi	DS1	K feldspar-schist,	S64/435134	5·58	546	383	0·60	22·8	2·7±0·1
			Tuke River			843	402	0·47	26·6	2·1±0·1
	Pl				1·30	383	544	0·56	45·6	11·1±0·1
						782	875	0·64	66·2	12·7±0.1
7561	Bi	DS2	Garnet schist,	S64/457100	3·44	881	514	0·57	42·5	4·3±0·1
			Tuke River			170	345	0·66	14·3	5·0±0·1
						749	448	0·47	34·1	3·5±0·1
	Pl				0·422	191	383	0·13	22·8	7·9±0·1
	Tr				1·30	368	389	0·22	24·1	4·4±0·1
7562	Bi	DS3	Biotite schist,	S64/478047	4·10	?	?	8·00	68·9	49·5±0·5
			Ivory Glacier			137	679	7·66	56·5	47·4±0·4
	Pl				0·595	216	882	1·08	66·1	46·0±0·3
	Tr				2·00	208	659	2·34	55·2	29·8±0·2
						760	1640	2·36	81·9	30·1±0·2
7563	Bi	DS4	Biotite schist,	S64/505121	3·50	818	542	0·71	45·4	5·2±0·1
			upper Cropp River			273	380	0·73	22·3	5·3±0·1
	Pl				0·131	82·6	355	0·06	16·8	12·4±0·1
						91·6	375	0·08	21·2	14·8±0·1
7564	Bi	DS5	Biotite schist,	S64/524116	4·30	498	1050	4·37	71·9	26·0±0·2
	Pl		Mid Cropp River		1·01	229	647	1·03	54·3	26·2±0·2
	Tr				3·16	294	518	1·60	42·9	12·9±0·1
						1220	1130	1·45	73·8	11·8±0·1
7565	Chl	DS6	Chlorite schist,	S64/512011	1·55	248	1250	3·98	76·3	64·9±0·5
			Seddon Creek/Wilkinson River			240	1240	4·08	76·2	66·4±0·5
	Pl				0·922	78·7	628	2·61	52·9	71·3±0·6

Notes: 1) Tr=total rock, Bi=biotite, Pl=plagioclase, Chl=chlorite

2) Decay constants K^{40} $=4·962 \times 10^{-10} yr^{-1}$

$=0·581 \times 10^{-10} yr^{-1}$

Isotopic abundance $^{40}K/K=0·01167$ Atomic %

total rock determinations are in close agreement at 4.3 to 4.4 Ma.

DS4 from the upper Cropp River 7.2 km from the Alpine Fault in the central low grade zone is a quartz, albite, biotite, chlorite, clinozoisite, calcite schist. Two schistosity producing events are noted from this rock separated by porphyroblast growth. Again the plagioclase ages are considered to be too high and the 5.2 to 5.3 Ma biotite ages are adopted.

DS5 from the mid Cropp River 8.4 km from the Alpine Fault is a garnet schist from the eastern garnet zone comprising quartz, albite, biotite, garnet, muscovite and calcite.

Biotite and plagioclase ages are in close agreement at 26.0 and 26.2 Ma and are considered to represent the last major cooling of these minerals. The 11.8 to 12.9 Ma total rock ages probably reflect the presence of a minor late mineral phase, probably chlorite.

DS3 is from the Ivory Glacier 10.4 km from the Alpine Fault in the eastern oligoclase zone. The rock contains quartz, oligoclase, biotite, chlorite and clinozoisite. Post-kinematic biotite porphyroblasts are developed over schistosity forming biotite plates. There may be late stage chlorite in the matrix. Garnet-bearing rocks outcrop nearby.

Close agreementbetween plagioclase and biotite ages give a 46 to 47.4 Ma cooling. Total rock age (29.8 to 30.1 Ma) may reflect a minor late stage mineral phase.

DS6 is from Seddon Creek in the Wilkinson Range 15.5 km from the Alpine Fault in eastern belt chlorite grade rocks. Composition is quartz, albite, chlorite, epidote, stilpnomelane. Original detrital grains and bedding are recognisable. Schistosity is delineated by chlorite growth.

The age range for this rock is 64.9 to 71.3 Ma. The older plagioclase age may reflect closure of the plagioclase system to argon migration at a higher temperature for plagioclase than for chlorite or possibly excess argon retained in plagioclase during low grade metamorphism.

Isochron Pattern

The map (Fig. 13) shows the positions of sample sites, the ages ascribed to them, and isochron contours interpolated from them. In all cases the ages considered reliable are from mica seperates. The feldspars and whole rock dates indicate incomplete degassing of argon at some stage in their history or alternatively a higher temperature of argon retention.

The isochron pattern while not considered precise in detail does show a consistent age trend. From a maximum of up to 70 Ma in the easternmost sample there is a gradual and consistent decrease in biotite ages to 26 Ma at 8.4 km from the Alpine Fault. Through the zone between DS5 and DS4 the age decreases from 26 Ma to 5.3 Ma and from there decrease in age becomes gradual to 4.3 then 2.7 Ma near the Alpine Fault.

This age pattern shows the expected age decrease toward the Alpine Fault as found by Sheppard et al. (1975) and Adams (1979). The one anomaly in this is the prominent two level division in the rate of age change toward the Alpine Fault. There is some sort of change between 7.2 and 8.4 km from the Alpine Fault which has produced the age pattern anomaly.

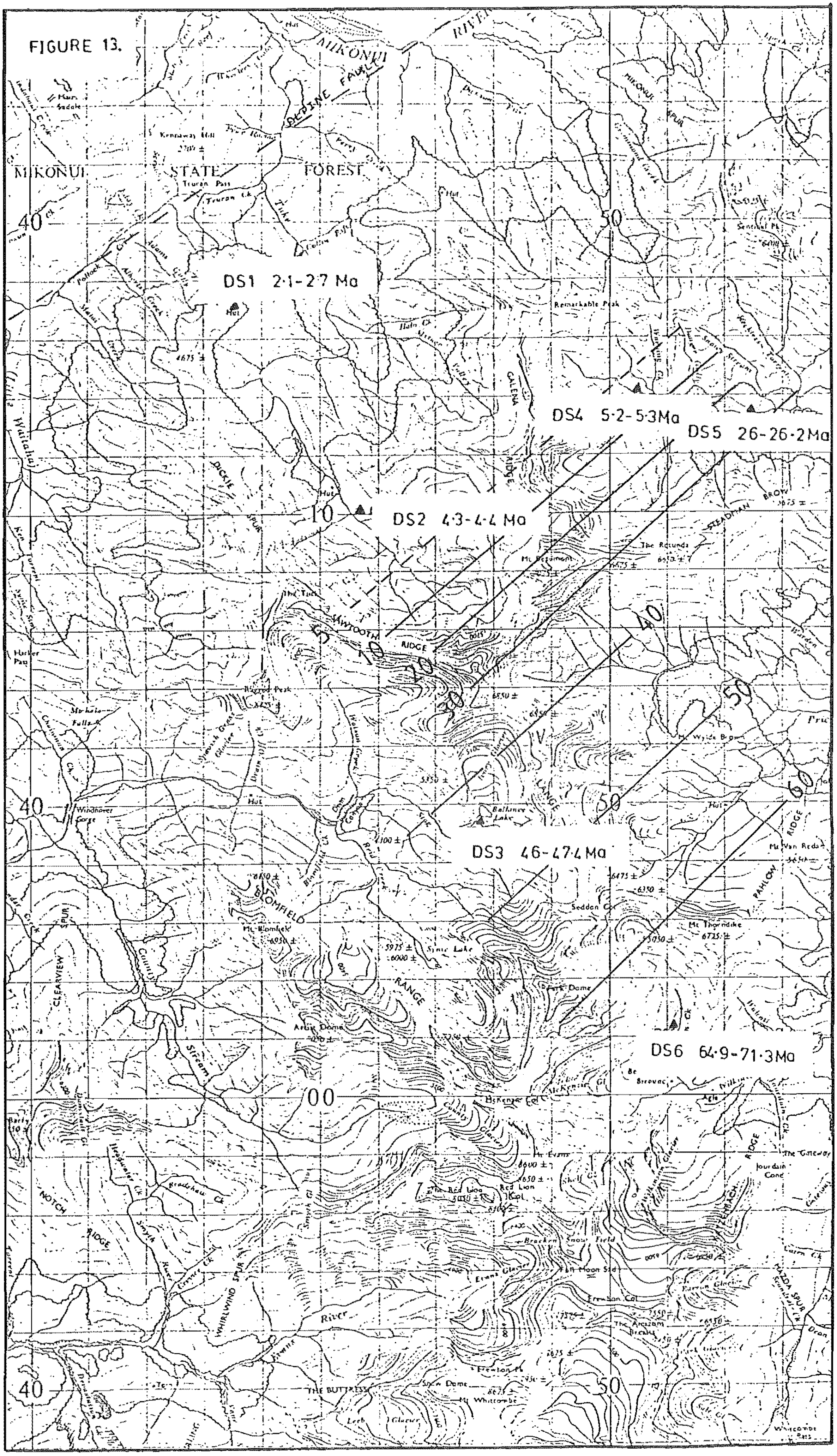
Given that the dates are measures of the last cooling of these rocks below the argon retention temperature ($200^{o}C$ to $250^{o}C$) there are several possible mechanisms for producing this pattern:

1 Differential uplift; the uplift of these rocks carried on to much more recent times in the northwest than in the southeast. This requires that the zone between DS4 and DS5 is a zone of vertical shear.
2 Reheating of the rocks in the northwest above the argon retention temperature in the period 26 to 5.3 Ma and the age increase from 26 to 70 Ma is some sort of aureole effect.
3 The changes observed are a relict of sample distribution, of inaccuracies in the dating procedure or of selection acceptance of data.
4 Some combination of 1 and 2 (and possibly 3) with reheating and differential uplift.

These alternatives are discussed in the light of structural and metamorphic data in section 3.6.

Fig 13:Map showing K Ar dates and isochrons for the study area.

FIGURE 13.


DS1 2·1-2·7 Ma
DS2 4·3-4·4 Ma
DS3 46-47·4 Ma
DS4 5·2-5·3Ma
DS5 26-26·2Ma
DS6 64·9-71·3Ma
MIKONUI STATE FOREST
ALPINE FAULT
MIKONUI
RIVER
SAWTOOTH RIDGE
BLOMFIELD RANGE
CLEARVIEW SPUR
NOTCH RIDGE
WHIRLWIND SPUR
PAHLOW RIDGE
STEADMAN BROW
GALENA RIDGE
DICKIE SPUR
Remarkable Peak
Mt Evans
Red Lion Col
Bracken Snow Field
Mt Whitcombe
Mt Thorndike
Seddon Col
Ballance Lake
Mt Blomfield
THE BUTTRESS
40
50
10
00

3.6 METAMORPHIC HISTORY

In a southeast-northwest transverse across the Alpine schists in the Cropp area the sequence encountered would be as follows:

1 Texturally low grade chlorite-biotite grade schists with a single schistosity. Evidence for one kinematic event and a coincident thermal metamorphic event.

2 Texturally low grade biotite, garnet and oligoclase grade rocks displaying a single schistosity and with post-kinematic biotite and garnet porphyroblasts.

3 Texturally intermediate biotite grade rocks with two schistosities and an intervening porphyroblastic biotite phase. A first kinematic phase has been followed by a thermal then another kinematic event.

4 Texturally high grade garnet and oligoclase grade schists with syn- and post-kinematic porphyroblast growth.

5 Texturally high grade K-feldspar-bearing schist with evidence for a post-kinematic thermal event.

The existence of such variability suggests a metamorphic history involving both spatial and temporal variations in kinematic and temperature conditions. Two interpretations are possible to explain the observations.

Metamorphism could have occurred as part of a single event with locally developed fluctuations in thermal and kinematic conditions. From southeast to northwest the thermal maximum increases in intensity, at low grades it coincides with a kinematic phase and at higher grades post-dates the kinematic phase. In the central zone (second biotite zone) there were two thermal maxima: one post-dating a kinematic event and a later one synchronous with second schistosity development. In this central zone either temperatures were never as high as in surrounding zones or retrogression accompanied the second event. Further west the kinematic maximum was long-lived and possibly involved pulses in places coinciding with porphyroblast growth, in others pre-dating growth. In the far west the late stage thermal maximum may have reached very high temperatures. Under this interpretation the age pattern discussed in section 3.5 is a product of post-metamorphic uplift.

Alternatively there could have been two high grade metamorphic events. During the first event oligoclase grade developed over a wide area. The kinematic maximum pre-dated the thermal maximum except in the lowest grade rocks and temperatures increased very slightly toward the west. The second event covered only half the area, it produced retrogression from garnet-oligoclase grade to biotite grade in the central zone (Fig. 14) with development of a new fabric coincident with the temperature maximum. To the west the second event obliterated evidence of the first (except for preservation of porphyroblasts) with the kinematic maximum pre-dating or accompanying the temperature maximum. Higher grade rocks are associated with this event.

The two alternatives differ only in the time break represented by the cooling before the development of the local second schistosity (area 3, Fig. 14) and on the inferred time difference between metamorphic maxima in the northwest and southeast.

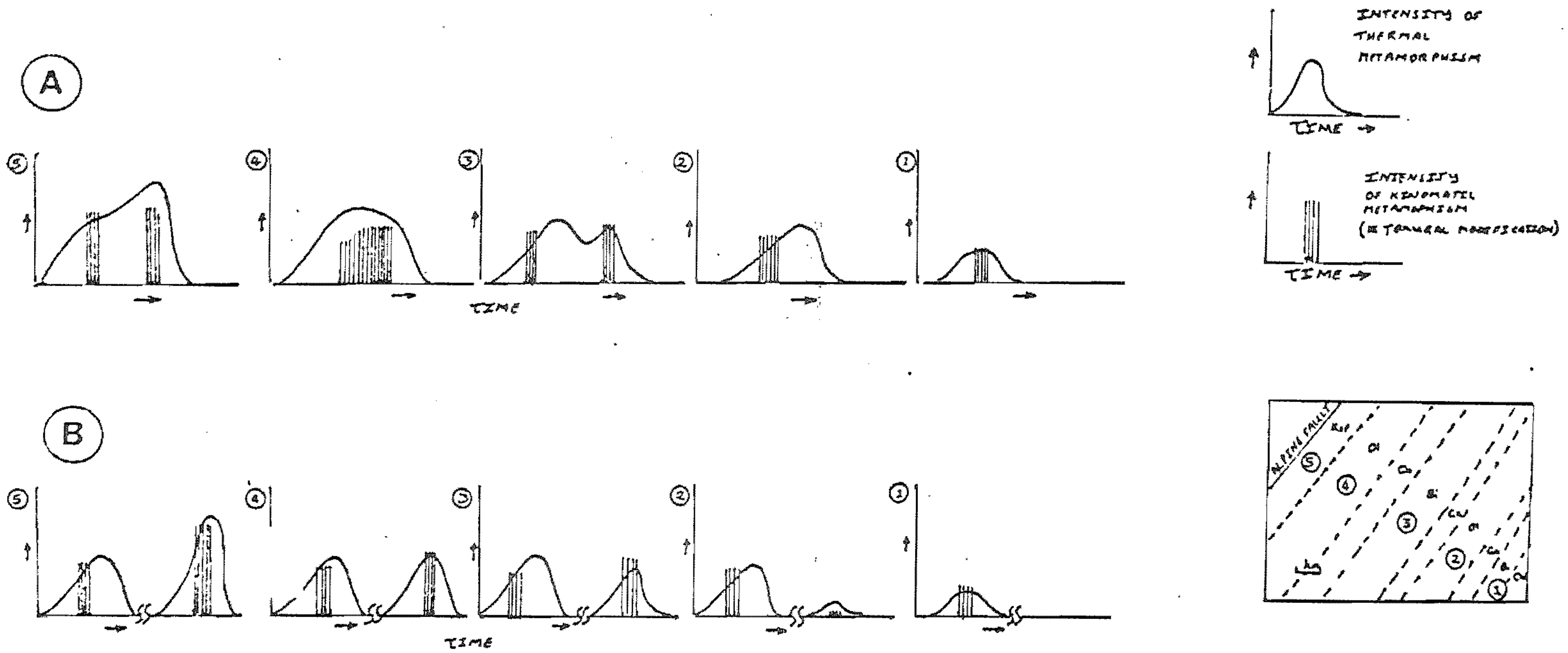
Textural development of the schists gives a clue to which of these alternatives should be adopted. Through the eastern garnet and oligoclase zones textural development never gets to the point of producing significant mineral segregation laminae. In contrast to the northwest mineral segregation laminae develop in the upper biotite zone and are well developed in the oligoclase zone.

The cooling age pattern also shows a significant two-level pattern. Texturally low grade rocks of the southeast all give old ages (26 to 70 Ma). The texturally high grade rocks of the central and northwest zones give young ages (5.3 to 2.3 Ma). The boundary between these age zones coincides, as far as has been determined, with textural and mineralogical changes suggestive of the incoming of a second metamorphic event.

Thus textural and isotopic evidence supports the view that the microstructural and mineralogical differences between schists in the northwest and southeast are due to two events of different character and age.

The two metamorphic event hypothesis is considered more probable as it can explain all the observed features. The first major metamorphic event produced a very wide band of oligoclase grade schist with a low degree of textural development. As mentioned in section 3.3 this lack of textural development could have been for a variety of reasons but probably relates to vertical schistosity and bedding planes allowing sufficiently rapid

Fig 14:Schematic showing possible interpretations of metamorphism in the Cropp area.A-one multiphase event of different magnitudes in different parts of the area.B-two events of different magnitude separated by a significant time gap.


A
B
TIME
TIME
INTENSITY OF
THERMAL
METAMORPHISM
TIME
INTENSITY
OF KINEMATIC
METAMORPHISM
TIME
ALPINE FAULT

exchange of metamorphic fluids that segregation banding was not developed.

This early metamorphic event is dated at 70+Ma. The younger dates in the southeast schist belt (45 and 26 Ma) show some combination of retrogressive and uplift effects.

The second metamorphic event started between 26 and 5.3 Ma (dates straddling the first appearance of the second schistosity). The event involved a very high temperature metamorphism near the Alpine Fault with decreasing T and P to the southeast. The central biotite grade rocks were retrogressed from pre-existing oligoclase grade with development of the new fabric. The southeastern oligoclase and garnet grade rocks may have been involved in temperatures up to chlorite grade as some growth of late stage chlorite is present and total rock ages have been affected.

The metamorphic history is complicated further by local zones of high textural grade rocks apparently unrelated to either of the above events. In these zones an early mineral segregation banding parallel to bedding and folded about schistosity (S_2) raises the textural grade from a regional IIA value to IV. As discussed in section 3.3 these are similar in style and occurrence to zones of high textural grade rocks, found in nearby low grade greywackes, associated with the cores of large, early recumbent isoclines. The textures may have been developed by a physical sorting process in sub-metamorphic conditions producing quartz-rich bands and micaceous matrix rich bands enhanced by mimetic recrystalisation without the development of conspicuous dimensional preferred orientation to form a schistosity (see also Findlay 1980).

The sequence of metamorphic events observed in the Cropp area is considered to involve three episodes:

1 An early folding with development of a foliation (S_1) on bedding (S_0) by local mimetic recrysta sation under low grade conditions.
2 A widespread garnet-oligoclase grade metamorphism (M_2) at 70+ Ma with development of schistosity (S_2) followed by static growth of porphyroblasts.
3 A less widespread metamorphic event (M_3) involving retrogression of high grade M_2 schists in the centre of the schist belt and development of very high grade rocks in the northwest. This involved development of a new schistosity S_3 with syn- and post-kinematic porphyroblast growth.

3.7 CONDITIONS OF METAMORPHISM

The metamorphic sequence established in the previous section comprises three phases of movement with development of a planar fabric (S_1, S_2, S_3) and three metamorphic recrystallisations (M_1, M_2, M_3). These events are sporadically represented over the area with S_1 M_1 never preserved with S_3 M_3. The nature of the time gap between M_2 and S_3 is not well determined but is critical to the interpretation of geological setting.

S_1 M_1 develops a mineral segregation banding in the cores of major isoclines by mimetic recrystalisation localised on bedding planes. This is only locally developed and presumably occurred in sub-metamorphic conditions in the presence of abundant water.

S_2 M_2 involves high temperature and at least moderate pressure (amphibole facies) with development of a strong foliation (S_2), a widespread post-tectonic static recrystallisation but with no development of mineral segregation banding. Biotite, garnet and chlorite zones established during this event are narrow suggesting steeply dipping isograds. A wide oligoclase zone suggests that isograds became flat-lying in the high grade body of the Alpine schists.

An explanation for the lack of development of mineral segregation banding involving steeply dipping schistosity and bedding is discussed in section 3.3.

A model involving steeply dipping bedding and schistosity at the time of metamorphism removes the need for large scale post-metamorphic rotation of the Alpine schists (Wellman 1979 and section 5.1). Such a model is in accord with an explanation for linear metamorphic belts involving frictional heat generation in deep elongate shear zones (Scholz et al. 1978) as distinct from burial type metamorphism as seen in the Otago schists (Bishop 1972).

Thus a model for formation of the older Alpine schists by means of horizontal compression across a proto-Alpine Fault lineament, involving high strains, relatively high temperatures, moderate pressures and low P_{H_2O} is favoured for producing M_2. The structural implications of such a model are discussed in Chapter Five.

The S_3 M_3 events are better preserved and more obviously associated with the Alpine Fault. S_3 is similar to S_2 in that it is always a steeply dipping schistosity formed axial planar to mesoscopic and macroscopic steeply plunging folds. The schistosity surface is defined by growth of oriented biotites.

M_3 is rather different to M_2 in that:

1 it is in part regressive (the central biotite zone);
2 it has a wide range of mineral facies representing variable P-T conditions and;
3 there is little evidence of low grade rocks associated with it. It is more restricted in areal extent than M_2 and involves more rapid change in metamorphic conditions.

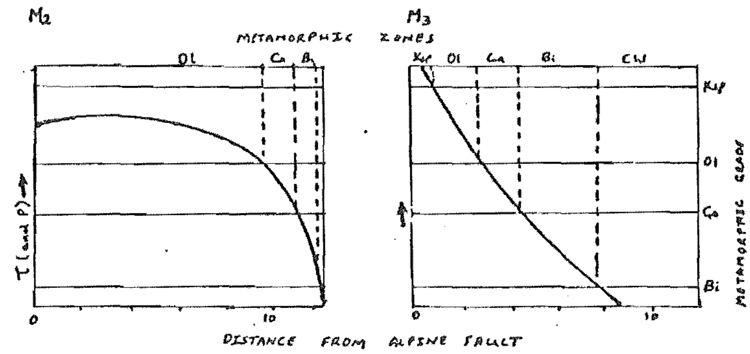
The pattern of M_3 metamorphic features is due to the superimposition of M_3 over high grade (M_2) metamorphic rocks. The M_3 involved only the western part of the M_2 metamorphic belt but involved a steeper geothermal gradient and higher temperatures than M_2. The regressive central biotite zone represents the area where M_3 temperatures were sufficiently high and H_2O availability was such that garnet (M_2) could be resorbed. At grades below this M_3 effects are restricted to minor chlorite growth (Fig. 15).

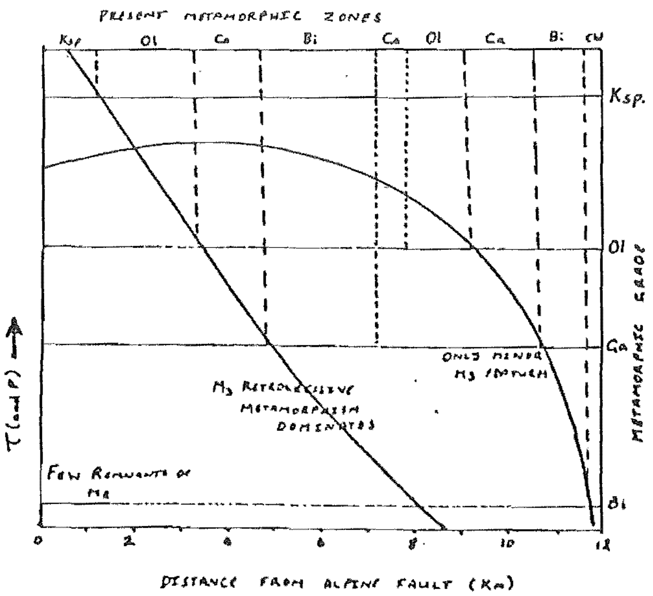
K-feldspar-bearing schists in the west of the area represent a M_3 temperature maximum. Findlay (1980) attributed minimum/maximum conditions of 450^{o}C at 3 Kba to similar rocks in the Franz Josef region. Winkler (1976) figures a curve for K-feldspar production from muscovite and quartz which gives 660^{o}C at 3 Kba. The texture of the K-feldspar is a web-like intergrowth with quartz and Na-rich oligoclase (Ab 85%). Both these features suggest high, possibly near anatectic, temperatures. Thus Winkler's value of 660^{o}C is considered more likely. Winkler also suggests that at P_{H_2O} values higher than 3.5 Kba at 660^{o}C anatexis will occur (Fig. 16) thus lower P_{H_2O} values are considered likely. This is in accord with the rock having been dehydrated during M_3.

The mineral assemblage may therefore represent conditions of very high temperatures with relatively low P_{H_2O} and P_{total}. Again the P_{H_2O} value is restricted because of the vertical rock fabric.

Scholz <u>et al.</u> (1978) presents a model involving frictional heating on the Alpine Fault to produce the heat required for Alpine schist metamorphism and found a favourable correlation between predicted heating and observed metamorphic effects. This model is compatible with either an entirely late Cenozoic or a Cretaceous and Cenozoic history of Alpine Fault movement.

Fig 15:Figure showing the different forms of the M_2 and M_3metamorphic events in terms of temperature and pressure distribution,and resultant metamorphic grades.


M2
M3
METAMORPHIC ZONES
Ol
Ga
Bi
Ksp
Ol
Ga
Bi
Cw
Ksp
Ol
Ga
Bi
METAMORPHIC GRADE
T (and P)
0
10
0
10
DISTANCE FROM ALPINE FAULT


PRESENT METAMORPHIC ZONES
Ksp
Ol
Ga
Bi
Ga
Ol
Ga
Bi
Cw
Ksp.
Ol
Ga
Bi
METAMORPHIC GRADE
T (and P)
ONLY MINOR M3 IMPRINT
M3 RETROGRESSIVE METAMORPHISM DOMINATES
FEW REMNANTS OF M2
0
2
4
6
8
10
12
DISTANCE FROM ALPINE FAULT (KM)

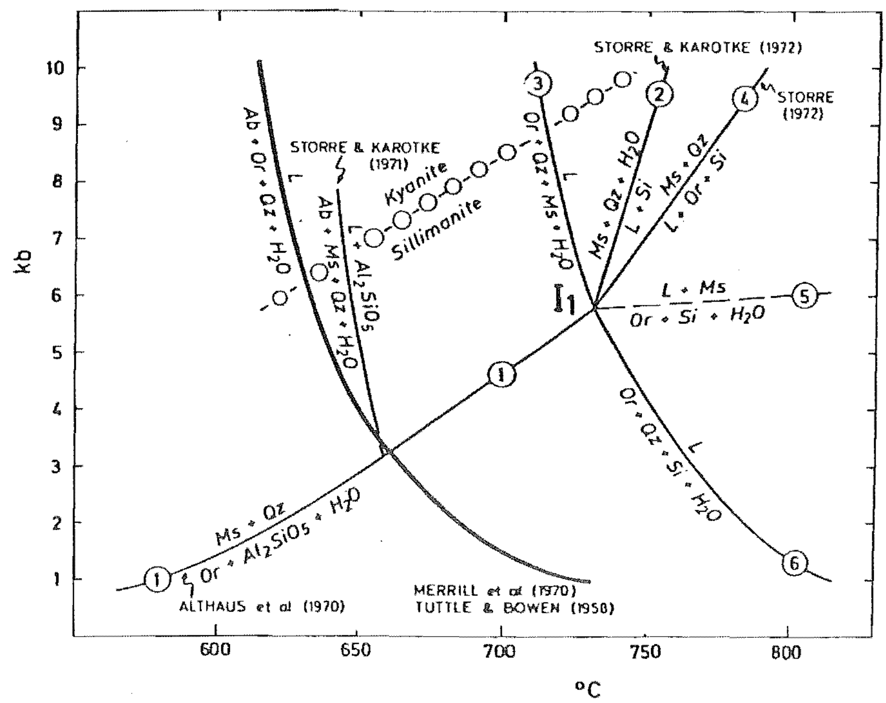


Fig.16:Reactions involving muscovite and quartz.Begining of melting in the system albite-K feldspar-quartz-watew also shown by heavy line.Or stands for K feldspar. (After Storre[1972],and Storre and Karatke [1971,1972])Extension of some of these reactions to 35kb has been studied by Huang and Wyllie(1973).From Winkler(1976).

Given that this model shows the availability of frictional heat to produce Alpine schist metamorphism there are certain problems with the Scholz et al. interpretation, because of a lack of basic mapping data from within the Alpine schists. Scholz et al. constrain the amount of Cenozoic uplift on the basis of the ratio of the widths of amphibolite and greenschist facies zones and on a lack of partial melting effects close to the fault. On the basis of the information presented in the previous section neither of these constraints is necessarily valid, the K-feldspar-bearing schists may reflect near anatectic conditions and the simple isograd pattern considered by Scholz et al. is probably not valid.

Fleitout and Foideveaux (1979), Brun and Cobbold (1979) and Nicholas et al. (1977) all describe metamorphism ascribed to heat generated by frictional heating in continental shear zones. All consider that conditions approaching anatectic temperatures can be established in shear zones similar to the Alpine Fault zone.

Present-day high heat flows in the schists near the Alpine Fault (Allis et al. 1979) and high geothermal gradients (90^{o}C/km measured at Franz Josef; Allis pers. comm., 1980) imply that metamorphic conditions exist in the upper lithosphere at present. This combined with an estimated 5 km of late Cenozoic uplift (C Adams, 1979) almost requires that schists exposed adjacent to the fault at present were metamorphosed in the late Tertiary as a result of Alpine Fault generated heat and with a steeply dipping fabric.

Thus the mechanism for producing the late Tertiary heating which caused outgassing of Ar^{40} as in Sheppard et al. (1975) can be extended to explain high grade metamorphism due to frictional heating both in the late Tertiary and in the Cretaceous.

CHAPTER FOUR: STRUCTURAL

4.1 STRUCTURAL SEQUENCE

Four deformation episodes are identified in the Cropp region, all involving folding (F_1 to F_4). The first three fold episodes are associated with metamorphic events (M_1 to M_3) and development of metamorphic fabric (S_1 to S_3). F_4 involves minor folding in response to fault movement.

F_1

F_1 folds are very large-scale and probably isoclinal. However, evidence as to the nature of F_1 is sparse. In the middle Cropp River area the S_2 schistosity cross-cuts small-scale isoclines (Fig. 17). F_1 folds are also indicated by a change in facing direction of F_2 mesoscopic folds between the Cropp River and the Ivory Glacier areas. This change in facing can be explained either by the presence of a large steeply inclined isoclinal fold or by a broad open east-west-trending steeply plunging pre-S_2 fold (Fig. 18). The isoclinal model is more compatible with isoclinal minor folds and with early structures observed elsewhere (Andrews et al., 1973).

As discussed in the previous section F_1 folding was accompanied by a low grade metamorphic recrystalisation (M_1) and production of a S_1 fabric parallel to S_0. This M_1 recrystalisation involved textural modification and development of textural zone IV rocks near F_1 fold hinges. These textures are preserved in the mid Cropp Valley and near Seddon Col (Fig. 19).

F_2

F_2 folds are associated with the dominant schistosity through the eastern half of the schist belt. Mesoscopic fold style varies depending on the degree of development of S_1. Macroscopic F_2 folds are steeply dipping, steeply plunging and tight to isoclinal (Fig. 20 and Map 4).

Through most of the lower Cropp to Ivory Glacier area mesoscopic folds are steeply plunging sub-vertical isoclines. Bedding is sub-parallel to S_2 on the fold limbs because of the tightness of the folds, but on a large-scale bedding is oblique to S_2 (see Map 4). Graded bedding is preserved in most of these rocks and younging directions can be determined although metamorphic grades are up to oligoclase zone over most of the F_2 dominated area.

Figure 17: F_1 isoclines refolded by F_2 about the S_2 axial cleavage.

a

b

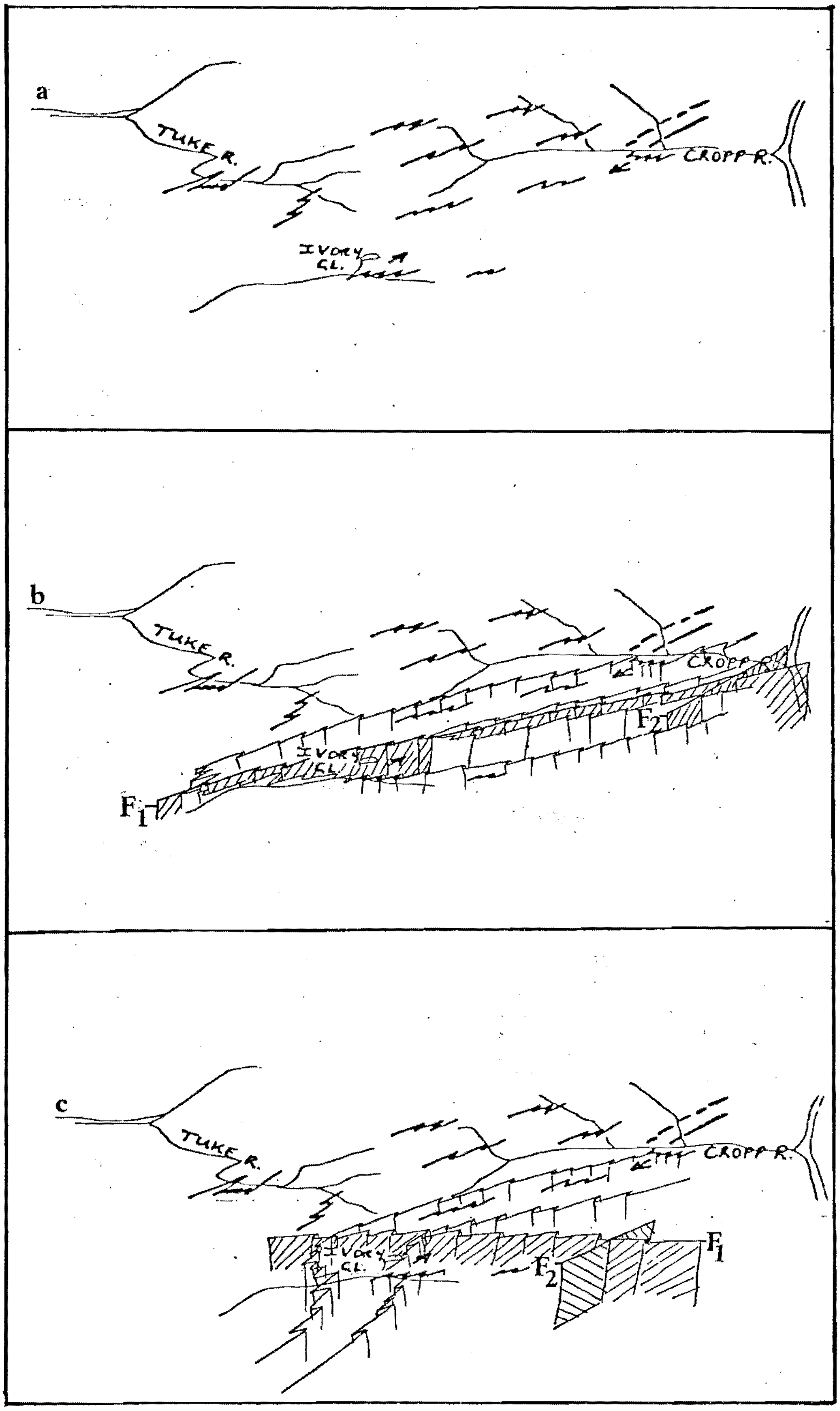
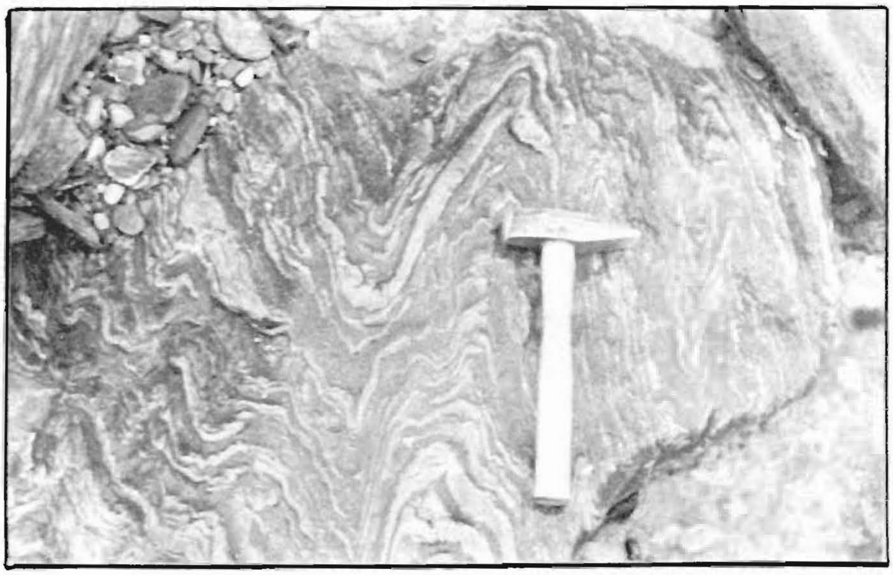


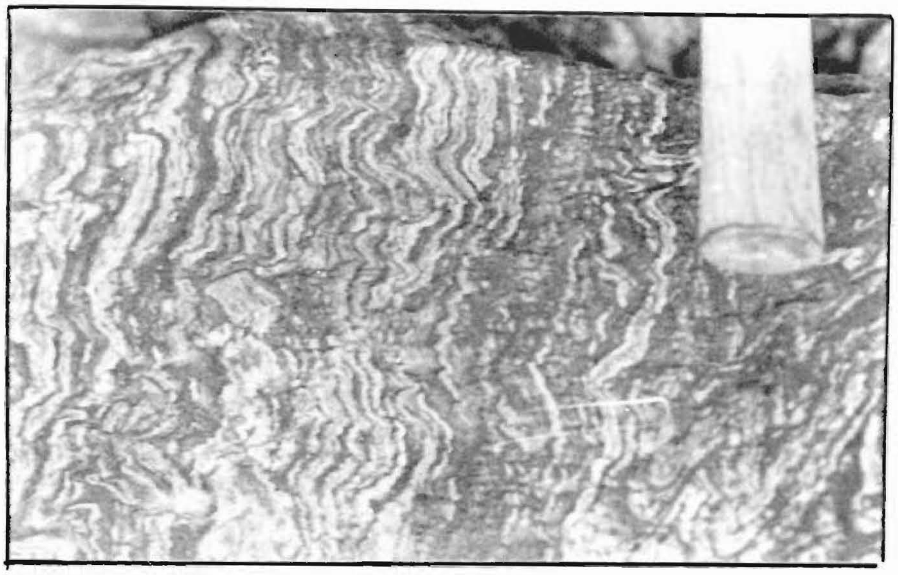
Fig 18: Evidence for large scale F_1 folding is given in (a), changes in mesoscopic F_2 fold faceing directions and vergance between the Ivory Glacier and the Cropp River. These can be explained by large F_1 isoclines (b) or broad F_1 folding cross cut by F_2 (c). solution (b) is preferred.

Figure 19: Mineral segregation laminae developed parallel to bedding during M_1 folded by F_2 and cut by S_2.

a



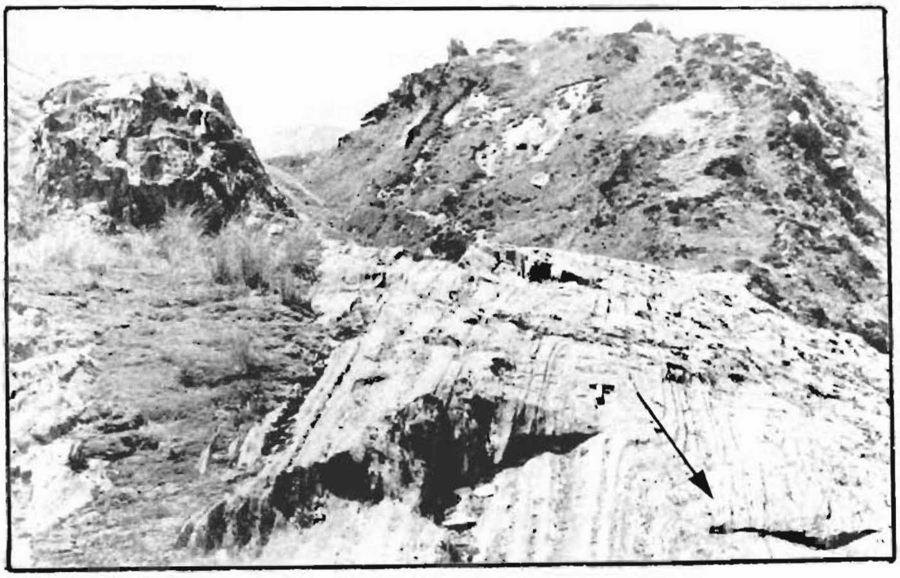
b



c

Figure 20: F_2 folds. (a) F_2 monocline in garnet grade quartzofeldspathic schist surrounding Pounamu Ultramafics near Noisey Creek. (b) F_2 isoclines with near-vertical fold axes typical of folds in quartzofeldspathic schist of the lower Cropp. From the Cropp River at the junction with Reckless Torrent.

a



b

In the mid Cropp where S_1 is developed, folds similar to Grindley's (1963) Otoko style folds (Fig. 21) are developed where S_2 transposes F_1 structures.

Emplacement of the Pounamu ultramafics preceded F_2 as S_2 structures are common in them (Fig. 22). They may have been emplaced in the cores of F_2 anticlines (Cooper 1977) but folding of the ultramafics by structures similar to F_2 to the northeast of the study area (Koons 1978) suggests that emplacement pre-dated F_2.

In the low grade rocks of the Wilkinson River F_2 folds are open, strongly assymetric and southwest-plunging with a steeply dipping axial fabric. These folds die out toward the non-schistose rocks of the (F_3) Mt Evans synform.

K/Ar dates on rocks dominated by the F_2 event are "mid" Cenozoic-late Cretaceous and therefore uplift accompanying or following the M_2 event was initiated before the late Cretaceous.

$\underline{F_3}$

The F_3 deformation event is almost coaxial with F_2 and is therefore not identified by distinctive fold styles. F_3 folds have an axial planar schistosity S_3 which is seen to displace F_2 features in the upper Cropp-Tuke and Ivory Glacier areas. F_3 is associated with a high grade metamorphic event in the northwestern part of the schist belt and retrogressive metamorphism in the central schist belt.

In the upper Cropp "Otoko" style folds are developed with S_3 transposing S_2. Further west "Paringa" (Fig. 22) style folds occur with S_3 completely dominating over S_2 structures. Because S_2 and S_3 are nearly parallel it is difficult to distinguish F_3 structures from F_2 dominated structures and the nature of the form surface of F_3 folds cannot be determined.

In the area showing Otoko-style deformation quartz and feldspar-rich bedding features are evident as rootless fold hinges. Within these beds S_2 is preserved and has been folded during formation of the rootless F_3 folds of S_0. Macroscopic structures delineated by changes in vergence of these minor folds show a corresponding change in the direction of convergence of S_2 and S_1/S_0 (Fig. 23). This suggests that the large-scale folds are F_2 structures which have been further compressed during F_3 to form tight rootless minor folds on the macroscopic fold limbs.

Figure 21: Paringa-style folds caused by the superposition of F_2 and S_2 on F_1. See Fig. 21c overleaf.

a

b

Fig 21:Different fold styles produced by intersection of S surfaces with different degrees of transposition involved in schistosity formation.From Grindley 1963.

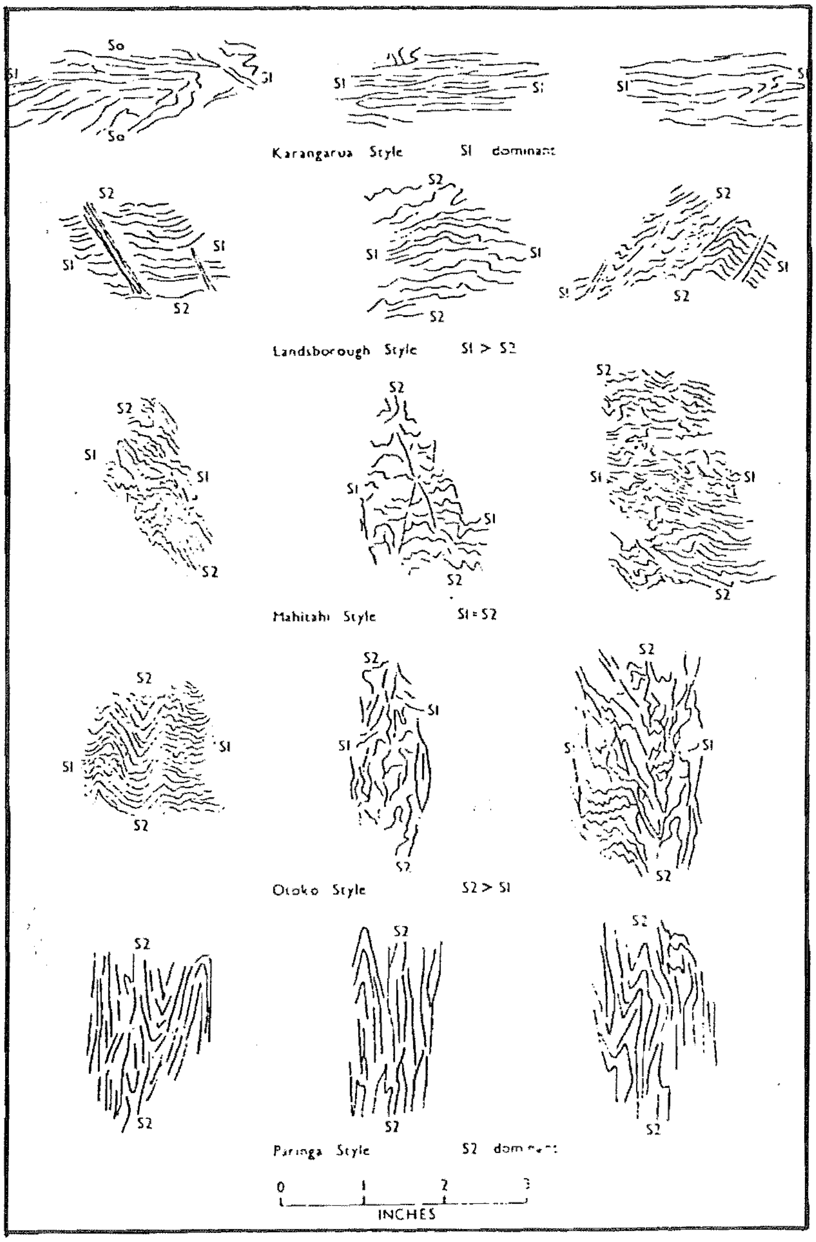

Karangarua Style S1 dominant
Landsborough Style S1 > S2
Mahitahi Style S1 = S2
Otoko Style S2 > S1
Paringa Style S2 dominant
0 1 2 3
INCHES

Figure 22: F_3 folds (a,b) Otoko-style rootless F_3 folds formed where S_2 dominates over S_2. S_2 structures are partly preserved in fold hinge regions (see Fig. 23); (c,d) F_3 folds Tuke River area.

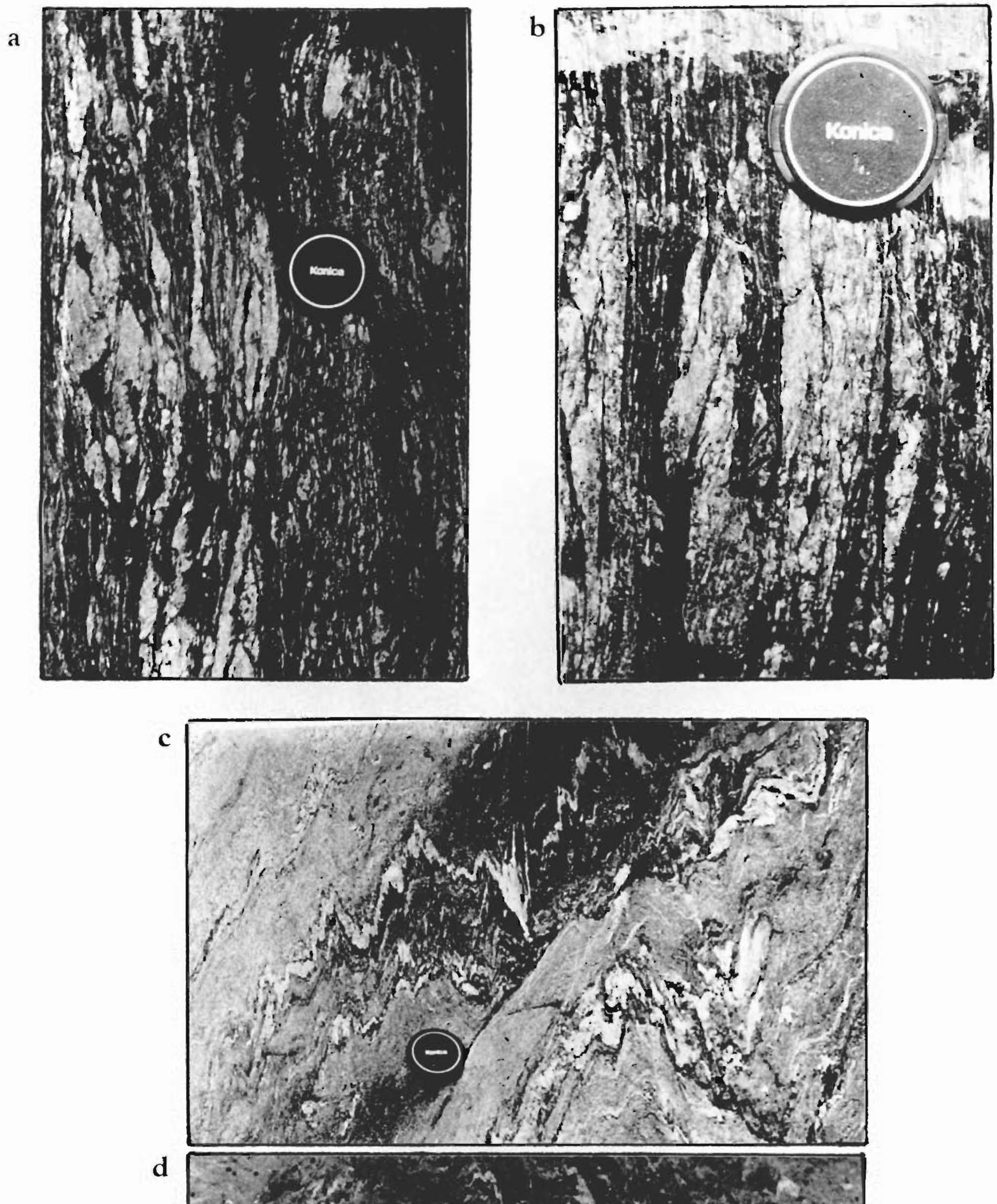
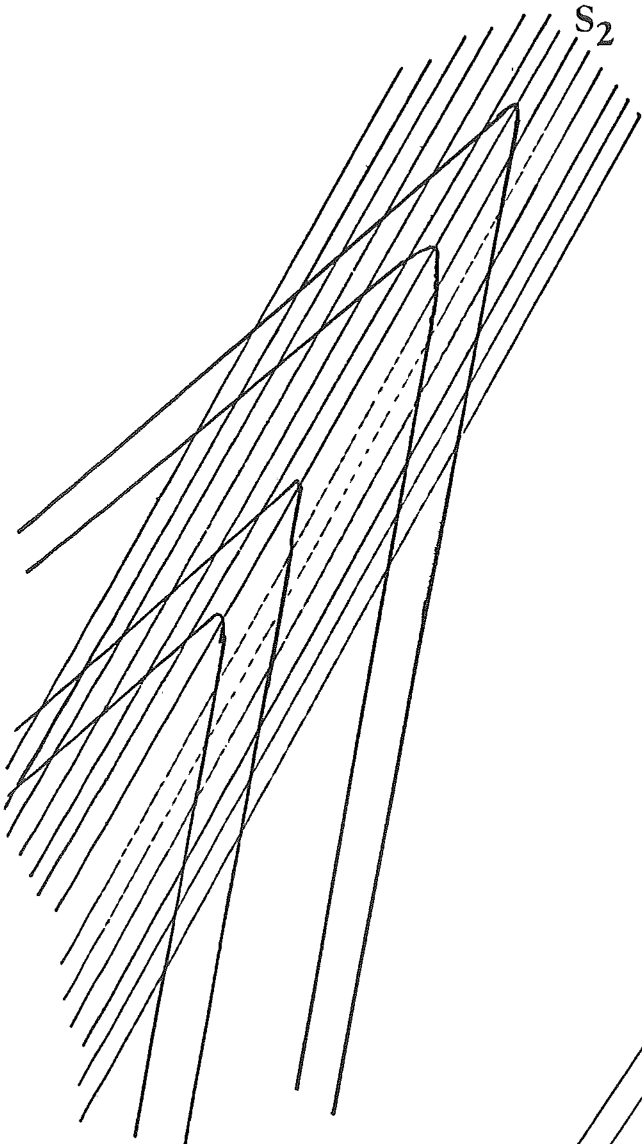
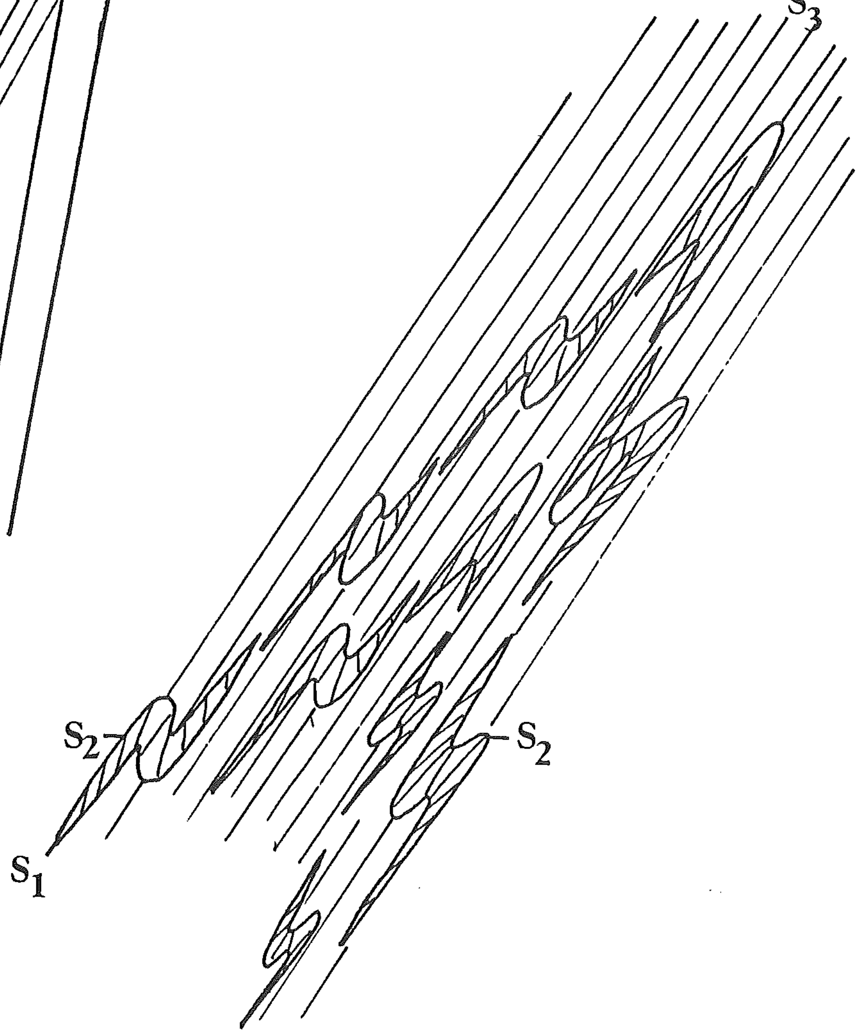

a
b
c
d
Konica

Fig 23: Rootless folds formed by overtightening and slip along cleavage when F_2 folds are subjected to coaxial F_3 folding. The S_2 fabric is preserved in the more competent beds.


S2


S3
S2
S2
S1

The area affected by F_3 folding gives K/Ar dates of 5.2 to 2.4 Ma. Uplift accompanying or post-dating the F_3, M_3 event was late Cenozoic.

F_4

Associated with a conspicuous ENE-trending steeply dipping, dextral fault set are moderately northwest-plunging small-scale kink folds. These folds are post-metamorphic, they involve complex brittle deformation and kink folding with no axial fabric (Fig. 24).

One of these faults displaced a marble bed 25 m+ dextrally in Steep Creek. There is no evidence for greater displacement on any of these faults. The Mungo Fault (Morgan 1908) may be a continuation of part of this fault set. A recent fault trace crossing Dickie Spur may indicate recent activity on faults of this orientation.

Probably of the same age as the ENE-trending faults is a set of sinistral north-trending steeply-dipping faults. The two sets are possibly part of some sort of conjugate fault set.

Other Faults

The Lake Lyes Fault is a major lineament trending NNE across the eastern boundary of the study area. It forms a set of parallel fault scarps across the east side of the Whitcombe Valley and forms the depression in which Lake Lyes lies.

It is offset in the north by the Mungo Fault (Cooper 1981, pers. comm.). It may continue to the south through the Price Basin and join a fault which passes through McKenzie Col.

This fault follows the eastern boundary of the Alpine schists as mapped by Morgan (1908). It is associated with high textural grade schists in the Frews Saddle area (A F Cooper, pers. comm. 1981).

The association with schistose fault rocks is similar to two faults mapped in the low grade greywackes to the southeast by Hawkes *et al.* (1980) and Bromell *et al.* (1980). All are apparently gently westward-dipping NNE-trending faults associated with schistose fault rocks and dividing significantly different structural, and possibly sedimentological, domains (Hawkes *et al.*, 1980). These may have been fundamental fractures during

Figure 24: Small (4cm) kink style F_4 folds from fault zone in Limpopo River.

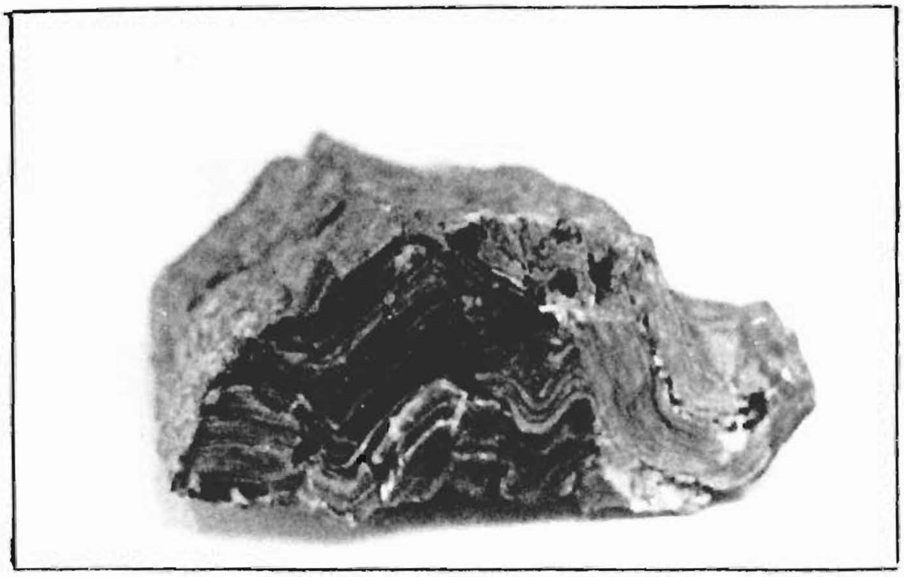
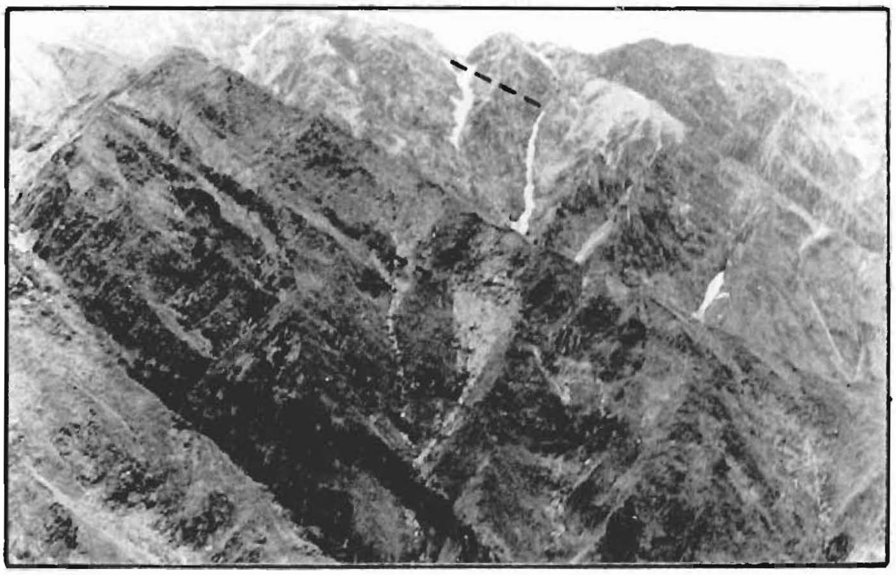


Figure 25: Late stage southwest dipping joint set controlling slope facets. Northern side of the Cropp catchment.



or immediately following F_3 during late Cenozoic uplift of the Southern Alps.

A late stage gently-dipping joint set is prominent on the northern side of the Cropp Valley. There is kinking of schistosity planes across these regularly spaced joints with movement generally down the joint dip. These joints are a late stage feature related to gravity collapse, but they are also a major control on geomorphology in the Cropp Basin (Fig. 25).

4.2 TIMING AND CORRELATION OF STRUCTURAL EVENTS

The preferred correlations of structural events described in this thesis with other works in the Alpine Schists is outlined in Table 4. Correlation of my F_4 event with events described elsewhere requires no explanation because of the similarities in trend and type of deformation. Other correlations are not so obvious. The reasoning behind these correlations and the inferred timing of structural and metamorphic events is explained below.

F_1 (this thesis) involves development of large isoclinal folds with incipient metamorphism associated with fold hinges. Similar large-scale folding is recognised throughout the Southern Alps as an early deformation phase. In South Westland this deformation phase is associated with metamorphism to greenschist facies (Cooper 1974).

In places a pre-F_1 folding phase involving large-scale folding of soft sediments has been inferred (Findlay 1980; Sporli 1975) but these are not recognised in the Cropp area.

F_1 folding is considered to represent the major deformation of the Torlesse clastic wedge, which occurred as a result of the collision of the allochthonous Torlesse wedge with volcanic arc rocks of the Caples terrain during the early phase of the Rangitata Orogeny (Bradshaw et al., 1980). In the Otago area this collision resulted in considerable tectonic thickening of the sediment pile, sufficient to produce high grade burial metamorphism. Toward the north, away from the collision zone, tectonic thickening and hence metamorphism was considerably less. Lower Jurassic K/Ar dates in Otago and the Chatham Islands (Adams and Robinson 1977) put a minimum age on this deformation phase.

TABLE 4: CORRELATIONS OF STRUCTURAL EVENTS IN THE ALPINE AND HAASt SCHISTS

	THIS STUDY		KOONS (1978) Arahura River area	ANDREWS et al (1973) Lord Range		WARD & SPORLI (1979) Ben Ohau Range
F_1	Indirect evidence for large scale isoclinal folding. Limited textural modification associated with fold hinges		Pre metamorphic very large scale isoclines	Possible very large pre F_1 folds (Sporli 1974) Large pre metamorphic recumbant isoclines and tectonic slides with incipient fabric only.		Exceptionally large isocli folds, schistosity locally developed axial planar to folds
F_2	Widespread steeply inclined, steeply plunging isoclinal folding with well developed axial schistosity forming regional "alpine" schistosity and the main metamorphic event. K Ar dates suggest a pre-mid Cretacious age.	F_1	Widespread folding associated with m metamorphism and "alpine" schistosity.	Strongly assymetric upright folds associated with steeply dipping syn-metamorphic cleavage. Folding coincident with development of isograds and isotects.		Steeply dipping schistosity developed accompanying or following folding of rocks to their present position
F_3	Folding nearly coaxial to F_2 in western schist belt. Folding associated with probable late Cenozoic metamorphic event.	F_2	Localy developed small scale folds with NE trends and low plungesddeveloped only in the western side of the study area. Axial surface defined by chlorite growth.	Post metamorphic faulting and possibly broad open folding (Dan synform).		Post metamorphic faulting
F_4	Local small scale kink folding associated with ENE trending faults		East west faults with kink folds.	NE trending faults		

TABLE 4: continued

THIS STUDY		FINDLAY (1979) Mt Cook-Copland R		GRINDLEY (1963) Haast R		COOPER (1974) Haast R
	F_1	Rare pre lithification folds				
F_1	F_2	Large scale isoclinal folds associated with metamorphism and widespread development of schistosity	F_1	Large recumbant folds with S_1 axial plane schistosity associated with regional metamorphic event	F_1	Syn metamorphic folding producing recumbant isoclines of nappe dimensions
F_2	F_3	Large step folds the vertical limbs of which form zones of intense transposition, intimately associated with the steeply dipping "alpine" schistosity.	F_2	Isoclinal south plunging folds with steep axial plane schistosity S_2. Associated with a resurgence of metamorphic conditions	F_2	Later syn metamorphic folding producing recumbant isoclines
			F_3	Post metamorphic open upright folding	F_3	Post metamorphic folding producing open structures with steep axial planes.
F_3	F_4	Minor folding about steeply inclined axial planes locally developed in the far west only.				
F_4	F_5	Minor steeply plunging kinks associated with dominant east trending joint set.		Faulting with local development of small kink folds	F_4	Kink folds associated with east trending faults
					F_5	Kink folding associated with NE trending faults.

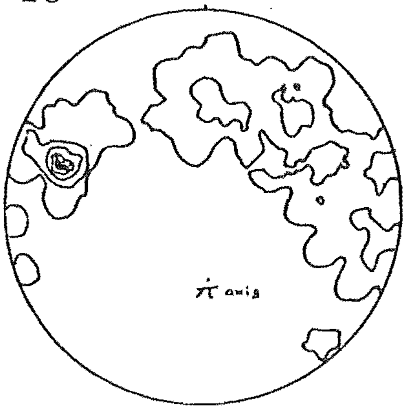
F_2 is the most prominent deformation event in the Cropp area and is associated with the most widespread metamorphic event. It involves isoclinal folds with a steep axial schistosity and is correlated with the major metamorphic events mapped in Central and South Westland. K/Ar dates constrain this event to 70 Ma+. As far south as the Haast River this deformation has a steeply dipping axial planar schistosity striking sub-parallel to the Alpine Fault. In the Haast River area Cooper (1974) mapped structures which are considered to be transitional between Otago and Alpine Schist type synmetamorphic structures. Cooper's F_2 folds are recumbent isoclines.

The dating of lamprophyre dyke emplacement in the Haast Valley is crucial to the interpretation of the structural sequence in South and North Westland. Descriptions of the lamprophyre dyke swarms in the literature are unclear as to how many episodes of dyke and sill emplacement there have been, or their relation to deformation phases. Published data on the dating of dykes in South Westland does not include the structural information necessary to show which intrusive/structural event is being dated (cf. Wellman and Cooper 1971 with Findlay 1980B).

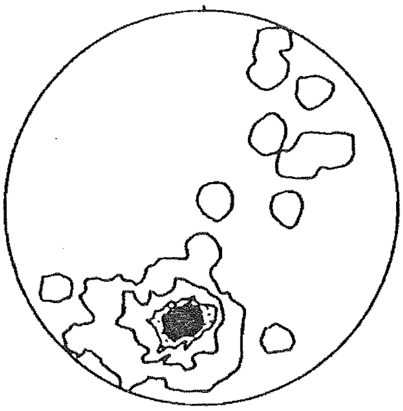
Cooper (1974, Fig. 26) shows plots of poles to dykes and sills at Haast which supposedly displays a point maximum corresponding to the maximum for F_2 fold axes and concludes that magma was therefore intruded along an a-c joint system. Alternatively the plot of poles to dykes and sills can be interpreted as a small circle distribution about a conical fold axis of 225^{0} 20 SW$\pm$. This is almost precisely on the F_3 fold axis maximum plotted by Cooper (Fig. 26e). As shown by Cooper the dykes post-date F_2 but dyke orientation appears to be affected by F_3. Thus three possibilities exist for the timing of dyke intrusion relative to F_3. Either the dykes were folded by F_3 or intruded into joints previously folded by F_3, alternatively there may be some complexity involving multiple intrusive phases which is not distinguished by the data as presented. The 100 Ma+ date on the "Haast" dykes accepted by Adams (1980) either puts a maximum age on the F_2 or the F_3 deformation and may date one of two intrusive events.

If Cooper (1974) F_3 pre-dates the 100 Ma dyke intrusion then it probably correlates with Grindley's (1963) F_2 and Findlay's (1979) F_3. In this case Cooper's F_1 and F_2 would correlate with Grindley's F_1 events or both Cooper's F_2 and F_3 (and Grindley's F_2) would correlate with Findlay's (1979) F_3 event (and F_2 this thesis). This last possibility is considered likely as Findlay's F_3 involves multiple deformation (R Findlay, pers. comm. 1981).

FIG 26

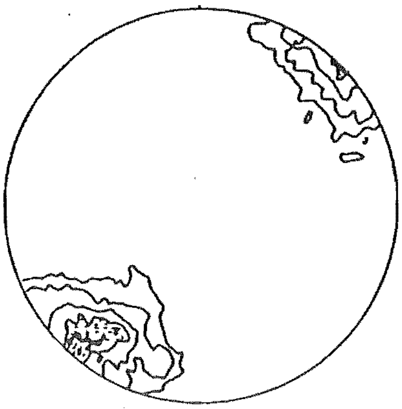


a. Poles to F_2 form surface



b. F_2 fold axes to which (e) is interpreted as representing poles to an a-c joint system.

c. Poles to F_3 form surface



d. F_3 fold axes.

e. Poles to post metamorphic intrusions in the Haast River area. Cooper 1973 interpreted the diagramme as representing point maxima from dykes and sills. The distribution of points is consistant with conical folding about F_3 fold axes and is what would be expected of initially steeply dipping dykes were folded about a sub horizontal cone axis. Diagrammes from Cooper 1973.

If the dykes are folded by Cooper's F_3 then there is a further possibility that correlatives of Cooper's F_3 are Findlay's (1979) F_4 and F_3 of this thesis and all are probably late Tertiary features. Either correlation is compatible with the rocks of the Cropp area and the alternatives can only be decided on the basis of rigorous mapping and dating of the lamprophyre dykes.

Everywhere in the Alpine Schists (with the possible exception of the Haast River area) the F_2 (this thesis) deformation produces a dominent steeply-dipping and northeast-striking schistosity near parallel to the Alpine Fault associated with K/Ar dates of mid-Cretaceous to mid-Tertiary. The younger ages are thought to be due to partial outgassing of Ar in the late Tertiary (Sheppard et al. 1975).

Rocks affected by F_3 are associated with late Tertiary cooling ages in the Cropp River. This suggests that deformation and metamorphism are also late Tertiary (26 to 5 Ma). Similar northeast-trending folding events in the western Alpine Schists have been noted by Koons (1978) F_2 and Findlay (1980) F_4.

F_3 folds and M_3 metamorphic rocks are only exposed near the Alpine Fault in a belt which appears to widen in the zone of greatest late Tertiary uplift (Fig. 27). In the north of the schist belt compressional movement has been taken up on the Hope, Clarence and Awatere Faults and in South Westland the compressional component of plate convergence (Walcott 1978) and hence uplift is not as great as in the Central Southern Alps and F_3 structures have not been exposed.

Other structures which may relate to F_3 deformation are the Dan-Mt Evans synform (see Map 4) and Cooper's (1974) F_3 folds discussed previously. Away from the schist belt F_3 deformation has been brittle with little significant folding of Mesozoic rocks since the earliest Tertiary (Oliver 1979), but with much faulting.

F_4 folding is associated with ENE-trending dextral strike-slip faults and north-trending sinistral faults. These represent a conjugate fault system developed near the surface accommodating the present-day compression across the zone.

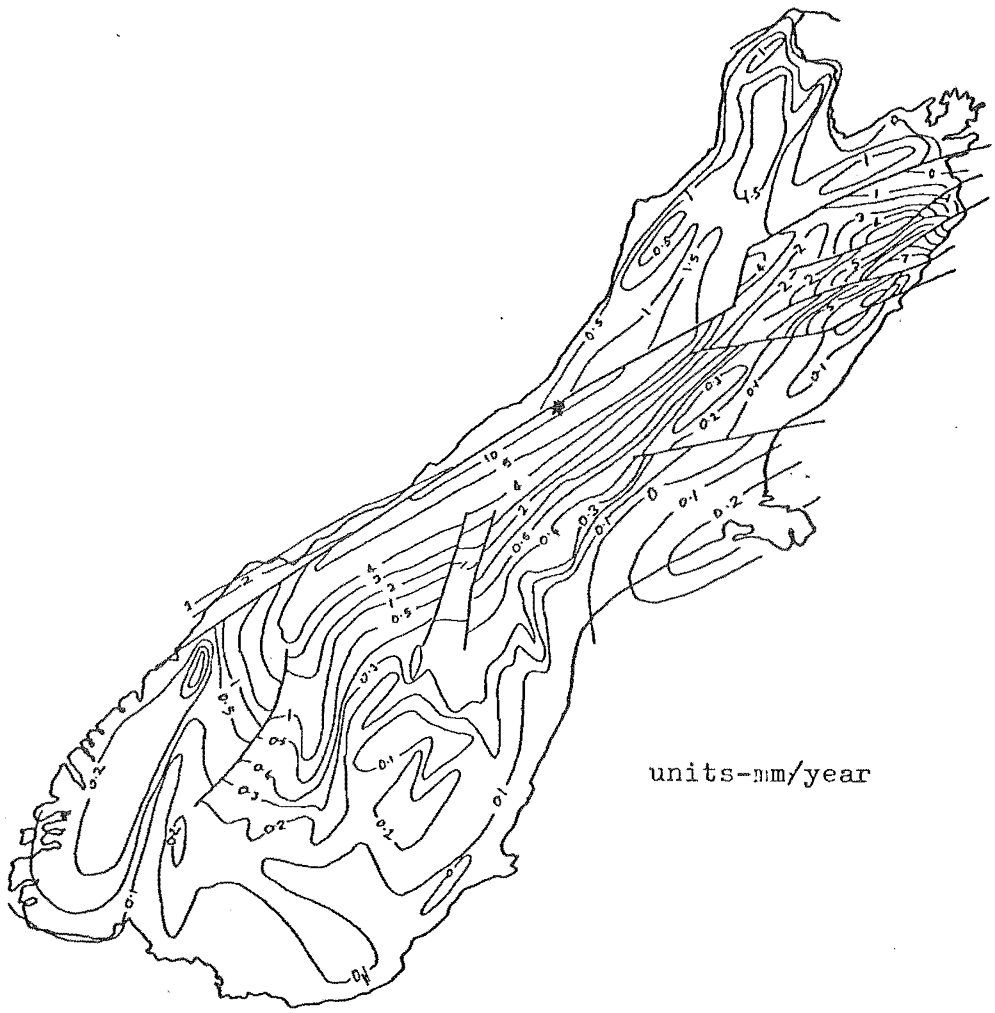


Fig.27:Map of South Island of New Zealand showing estimated uplift rates,from Wellman(1979).✱indicates study area.

CHAPTER FIVE: GEOTECTONIC SYNTHESIS

5.1 GEOTECTONIC MODEL

Any geotectonic model for evolution of the Alpine Schists must be compatible with the following five constraints:

1 The temperature range of argon outgassing ($250^oC \pm 50^oC$; C.Adams 1979).
2 The K/Ar date pattern especially the narrow band of young ages near the Alpine Fault.
3 The present high near-surface geothermal gradient near the Alpine Fault ($90^oC/km$; Allis, pers. comm. 1980).
4 The metamorphic grades attained in the Alpine Schists and the P-T conditions implied by them.
5 The steep dip on S_2 and S_3 surfaces, which parallel the Alpine Fault and the steep dips on the high grade isograd surfaces.

The model proposed conforms to these criteria. It involves four deformation phases all of which are represented in the Cropp area:

1 The earliest deformation recognised (F_1 and associated S_1 and M_1) involved tectonic thickening of a sediment wedge by formation of nappe-like folds as a result of collision of rock masses at a convergent plate margin. Metamorphic effects associated with this deformation are reduced away from the collision zone. The continental collision produced the Rangitata I orogenic episode of Bradshaw et al. (1980) and resulted in the major metamorphism of the Otago schists probably in the late Triassic (180 to 200 Ma).

2 A transcurrent fault developed along the line of the present Alpine Fault at a high angle to the trend of the Rangitata I orogen et 90 to 110 Ma (C Adams 1979b). Frictional heating in the clastic wedge overlying the faulted sialic crust produced a thermal antiform eventually resulting in the development of greenschist-amphibolite facies metamorphic conditions (Scholz et al. 1979). Uplift accompanied this event, the late stages of which saw a considerable compressional component across the schist belt. There may have been further deformation of the Otago schists at this time (producing Cooper's 1974, F_2 folding). In the low grade rocks away from the fault large-scale folds formed in what was possibly a wrench regime (Ward and Spörli 1979). The transcurrent part of this event occurred before the development of tensional basins on the West Coast and Central Otago at

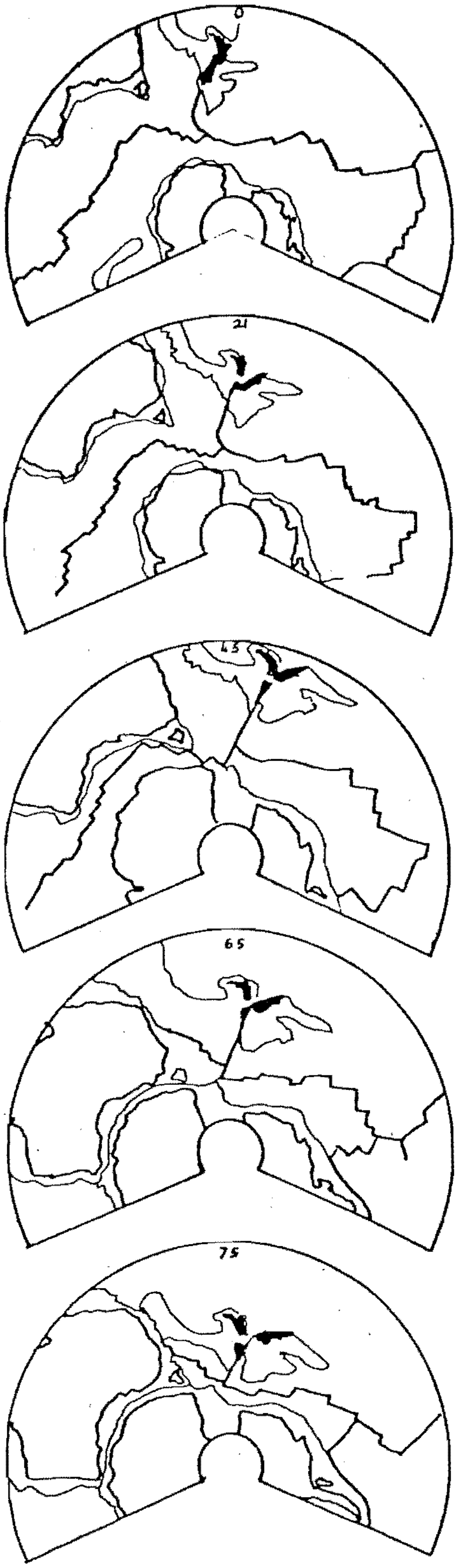
90 to 110 Ma (Adams and Nathan 1978) and the uplift ceased before 70 Ma (this study). Major uplift and transcurrent movement need not have been synchronous.

It is interesting to speculate as to the nature of the regional tectonic setting during the metamorphic and structural event. Although the Alpine Fault has displaced Paleozoic rocks 480 km plate tectonic reconstructions indicate that at least this amount of displacement must have occurred during the late Tertiary. This leaves no room for major displacement during the Jurassic-Cretaceous. Weissel et al. 1976 reconstruct the Southwest Pacific from sea floor spreading data from 80 Ma to the present. In their reconstruction at 75 to 75 Ma the Paleozoic belts are not aligned but are offset from their original position by a significant sinistral displacement (Fig. 28).suggesting a pre-75 Ma sinistral Alpine Fault. Molnar et al. 1975 accommodate this apparent anomaly by internal deformation within the New Zealand continent. They do, however, suggest that between 81 and 63 Ma fracture zones in the New Zealand region had sinistral offsets which ceased to be active at the time of initiation of Tasman Sea opening (63 to 35 Ma). Geological evidence for Cretaceous sinistral tectonics is not overwhelmingly obvious although Bradshaw (1977) indicates that the Triassic Esk Head mélange was displaced sinistrally across the Hope Fault which is now part of a dextral strike-slip fault system. Late Tertiary sea floor spreading data predicts an offset of 900 km across the New Zealand region since 38 Ma, 300 km more than the offset of the Dun Mt Ophiolite Belt (Walcott 1979). Some of this extra 300 km may have been produced by sinistral offset across the Alpine Fault during the Cretaceous.

Deformation and metamorphism associated with the shear system formed an Alpine Schist belt with originally steeply dipping schistosity (see Grindley 1963, p. 911) and mineral isograds, uplifted to within 12 km of the surface by the mid-Cretaceous (C Adams 1979).

3 Mid to late Tertiary dextral transcurrent movement on the Alpine Fault re-established the frictional heating regime close to the Fault and produced metamorphic temperatures at relatively shallow depths. A new steeply dipping schistosity developed near the fault and was involved in a dynamothermal metamorphic event probably under unusual pressure (P_{total} greater than P_{H_2O}) and temperature (section 3.7) conditions.

FIG 28



Reconstructions of the S.W. Pacific from sea floor spreading data(after Weissel et al 1976) Note that beyond 45Ma the north-western South Island has been displaced beyond the south-west South Island although the geology of the two areas suggests that they were originally adjacent.These reconstructions imply a pre 75Ma sinistral displacement along the trend of the Alpine Fault.

Oblique compression since 10 Ma has produced massive uplift adjacent to the Alpine Fault. This uplift has migrated southward with time. Uplift has been greatest in the area between the intersection of the Hope and Alpine Fault systems, south to the Mt Cook region. To the north compression is dispersed over a wider area and localised as movement on the branch faults. To the south uplift is lessened because of a shorter history of compression. As a result the metamorphic rocks related to these movements are only exposed in the Central Southern Alps although young K/Ar ages produced by partial degassing of older schists are widespread (Scholz et al. 1979).

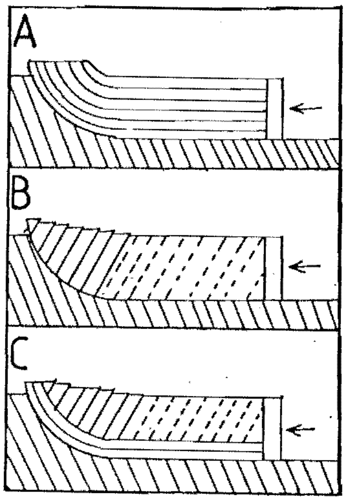
4 Near-surface post-metamorphic faulting sympathetic with compression across the schist belt is the most recent tectonic feature identified. Uplift, shear and metamorphism are still occurring at relatively shallow depths in what is a presently very active shear zone.

5.2 IMPLICATION OF THE MODEL TO THE UPLIFT HISTORY OF THE SOUTHERN ALPS

Tectonic models involving uplift of the Alpine schists belt from an original flat-lying position at the base of the crust up a curved fault plane (Wellman 1979) to explain uplift of the Alpine Schists conflict with geologic observations and do not explain the observed features as well as more conventional models. There are problems reconciling such a model with: steeply dipping schists of apparent late Tertiary age, Cretaceous uplift across a fault both sides of which comprised Torlesse Group sediments lying on oceanic crust, the structure of the Haast region where the supposed Alpine Schist conveyor belt intersects the Otago Schist megaculmination, and topology (see Fig. 29). A crustal model similar to Woodward (1979) is preferred with a zone of plastic deformation localised east of the Alpine Fault.

There is no need for very large, late Cenozoic uplifts of the Alpine Schists. K/Ar dates restrict the depth of burial of schists in the eastern part of the schist belt to the depth of the 250^{o}C isograd (at the time of closure of the mineral systems to argon loss) since the late Cretaceous. Assuming a geothermal gradient similar to Fig. 24 there is no need, from metamorphic or structural geological evidence, for uplifts greater than 15 km in the late Cenozoic (uplift rates of 4 to 6 mm per year if deformation is spaced over the last 3 Ma).

Fig 29:Evidence for rejecting the crustal model of Wellman(1979).



Three cross sections showing possible faulting patterns for formation of the Southern Alps,from Wellman (1979).Wellman considers that alternative C.is realistic with steeply dipping reverse faults there is a realistic uplift pattern ,and the schist is upturned.

Wellman's(1979) crustal modelis rejected because:

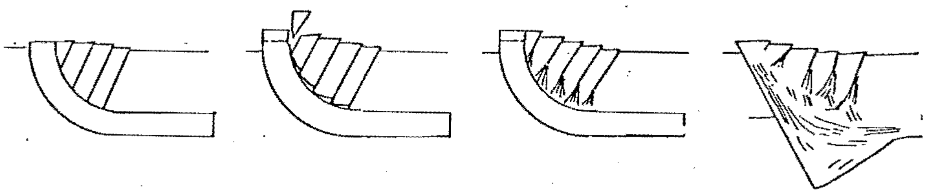
1) Model C.requires a near 90° rotation of alpine schistosity along the length of the Southern Alps.

2)In the Haast region the schist belt is 40km+ wide.90° block rotation of this thickness of material(greater than Wellman's assumed crustal thickness)would pose conspicuous topological problems.

3)Rotation predicts uplift at the surface greater than that observed.To uplift schist from the base of the crust(30km according to Wellman)to the surface up a curved fault plane would result in removal at the surface of material equivalent to the length of the pathway traveled(approximately $\frac{1}{2}\pi.30$km=47km).

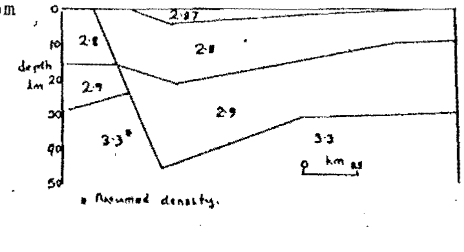
4)K Ar dates show that significant uplift occured during the Cretaceous .Therefore either a similar tectonic regieme existed during the Cretacious or all the rotation proposed by Wellman occured above the 250°C isotherm.

5)A dynamic interpretation of model C.predicts the developement of microfractures or possibly schistosity at the base of the reverse faults.



The crustal model of Woodward(1979)is preferred.Based on gravity data this model is compatible with a tectonic regieme involving compression and deformation across the Alpine Fault zone producing an initially steeply dipping axial schistosity

Crustal model from Woodward(1979).



Interpretation of K/Ar data requires a geothermal model for the Southern Alps different from that proposed by Allis et al. (1979). Allis' model is based on the "conveyor belt" uplift model which is rejected (see Fig. 29) and assumes a maximum temperature of 500°C for the schists at depth which is not necessarily justified especially if frictional heating is significant near the Alpine Fault (Scholz et al. 1978). The Allis and Scholz (frictional heating) models are presented in Fig. 30, together with a suggested model compatible with K/Ar and structural data.

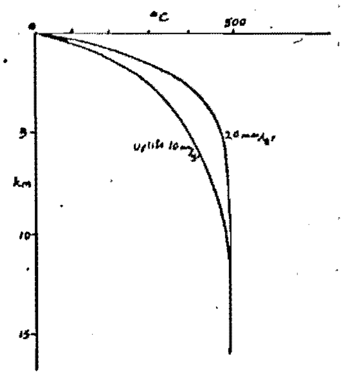
The most recent tectonic activity recognised is movement along ENE-trending faults. Such movement if active at present is incompatible with movement on "schistosity plane faults" (Adams 1979) being the major uplift mechanism in the schist belt. It is considered probable that vertical movements are occurring at depth by plastic deformation of schist near the Alpine Fault (slip folding) with sympathetic fault movements in lower temperature near-surface rocks. In view of the geothermal interpretation presented in Fig. 30c there is no geological requirement for strong differential uplift across the study area.

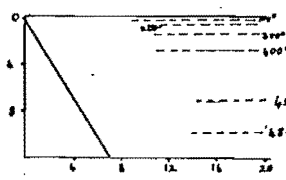
Long-term uplift of 20$\pm$ mm yr^{-1} in a belt parallel to the schist belt and involving the entire schist belt (J Adams 1978, p. 22) cannot be resolved in terms of the faulting patterns observed in the study area and elsewhere. Geomorphic considerations do not require long-term uplifts much greater than 10 mm yr^{-1} and suggest that uplift is occurring as block movement with no suggestion of drag effects adjacent to the Alpine Fault. (Evidence from within the study area and South Westland is discussed in section 6.5.)

Fig 30: Models of geothermal gradients under the Southern Alps.

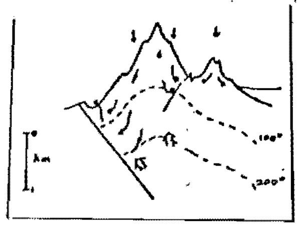
A. GEOTHERMAL MODEL: Allis *et al*(1979)

One dimensional model for zone 15-10km from fault.





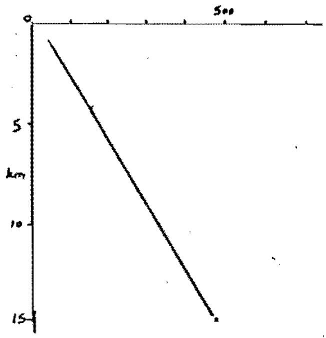
In two dimensions the model is unrealistic in that it predicts that the area 5-15km from the fault is isothermal with depth

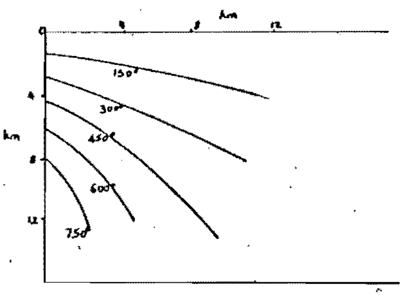


Allis et al(1979) fig 5.

B. GEOTHERMAL MODEL: Scholz *et al* (1980)

One dimensional model for zone 12km from fault.



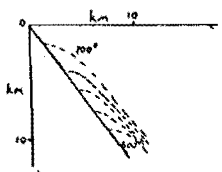


Two dimensional model is more realistic than A. in that gradients decrease away from the Alpine Fault.

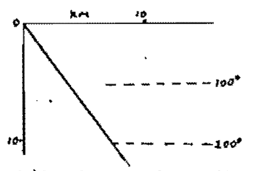
Although the two models are based on different time perspectives(one Cretaceous,the other Cenozoic)they interpret features of the same metamorphic belt and the same KAr data.

C. GEOTHERMAL GRADIENTS: Preferred construct.

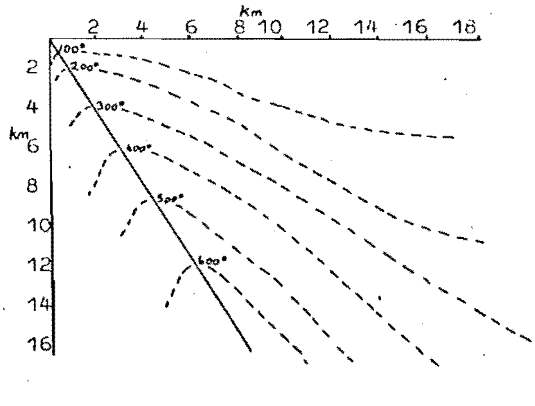
The preferred construct is based on Allis' estimates of very high geothermal gradients near the surface within two kilometres of the Alpine Fault (90-100°C/km)produced by a combination of upward transfer of heat and mass and by frictional heating in the fault zone.The temperature gradient decreases rapidly with depth near the Alpine Fault to 50°C/km The normal geothermal gradient away from the fault zone,the zone of frictional heating,and where thermal conductivity is lower than mass transfer rates is assumed to be 20°C/km.The model is also constrained by the presence of a steeply dipping fabric which inhibits lateral heat transfer.



a)heat scource on fault; no other heat input. uplift steepens geothermal gradient in top 2km,the zone of meteoric water circulation.



b)background geothermal gradient.



CHAPTER SIX: GEOMORPHOLOGY (I)

6.1 INTRODUCTION

The study area is in mountainous terrain and is subject to extreme precipitation; its geomorphology reflects a history of recent glaciation and strong fluvial downcutting.

The present river systems are very immature, having steep gorges, uneven stream profiles (including large waterfalls) and headwaters which recently have, and in some cases still are, influenced by glacial action.

Geomorphic history of the Cropp area is discussed in terms of controls on drainage patterns, identification and characterisation of glacial events, post-glacial adjustments and regional neotectonics. Relevance of the geomorphology to erosion studies is discussed in section II Chapter Eleven.

6.2 CONTROLS ON DRAINAGE PATTERNS

Drainage basin shape in the Cropp and Tuke Rivers is controlled by an interaction of bedrock features and run-off characteristics. A prominent west-striking fault set controls minor streams draining Galena ridge and the sides of the Tuke Valley. A second set of north-trending faults controls drainage on the northern side of Cropp Brow and in the gorge of Noisey Creek.

Resistent rock units (ie chert and ultramafics) cutting across drainage directions are the cause of many waterfalls and gorges (Fig. 31). The Pounamu ultramafics are a distinctive resistant rock unit. Striking along the contour they provide a barrier to normal drainage and valley deepening in Noisey Creek.

A moderately dipping joint set on the north side of the Cropp Valley controls south-facing slope angles (Fig. 25). This plane acts as a preferential zone of failure for gravity collapse possibly promoted near the surface by freeze-thaw movement. Disruption of bedrock associated with this joint set makes the rock highly permeable and may account for the rapid run-off characteristics of the Cropp catchment (discussed in Chapter 9).

Figure 31: Upper gorge in Noisey Creek. Glacial planed bedrock surfaces are preserved where fluvial downcutting has been slowed by the presence of resistant ultramafic rocks in the upper gorge of Noisey Creek.

6.3 GLACIAL FEA

Pleistocene

Ice planed surfaces resulting from glacial action are easily distinguished. Recent fluvial direction of the area has produced a rugged topography with slopes typically exceeding 30^{0} and often 55^{0} to 70^{0}. In sharp contrast to the fluvial dominated topography there are numerous slope facets, flat topped ridges and smoothed bedrock surfaces which grade to old valley levels above the present valley bottom which are considered to be remnant glacial planed surfaces.

The oldest set of glacial planed bedrock surfaces grades to the tops of the major ridges in the Cropp and Tuke Rivers area and corresponds to similar surfaces on the west side of the Main Divide in the mid-Whitcombe River valley.

In the Tuke River the valley floor graded from the ridge south of Mt Beaumont and off the slopes of Mt Beaumont near Hanging Valley Creek (Fig.1), across Flat Spur, truncated Dickie Spur on its north end and several spurs on the north side of the Tuke (Fig. 32). An elevation range of 2200 m to 1100 m. There may be two levels in this planation.

Comparable surfaces in the Cropp River grade from the flat topped Galena ridge (~2000 m) and off the upper part of the Steadman Brow. In the main Cropp Valley aligned ridge tops and slope facets delineate a valley floor at about 1000 m to 1400 m. Noisey Creek and Reckless Torrent held small feeder glaciers which flowed into the Cropp (see Map 5 and Fig. 33).

The ice cover at the time of formation of these surfaces was in the form of a broad, widespread ice field or snow dome with only a few peaks emergent above the ice (Sawtooth Ridge and the Lange Range). Most of the peaks in the area were ice-covered (Fig. 33d) and the valleys where ice flow was concentrated had a relatively subdued relief.

The high level ice planed surface of the Cropp area correlates with the last phase of glacial valley widening in the Whitcombe Valley and with the last surfaces suggesting glacier movement out of the alpine valleys and onto the Hokitika flood plain. Similarly in the Tuke there is no evidence to suggest that glaciers have extended beyond the mid-Tuke after this high level planation.

Figure 32: Glacial planed surfaces in the Tuke River (from Dickie Spur). Lightly stippled surfaces were developed during latest Pleistocene glaciation. Shaded surfaces represent areas affected by early Holocene glacial events.

a

b

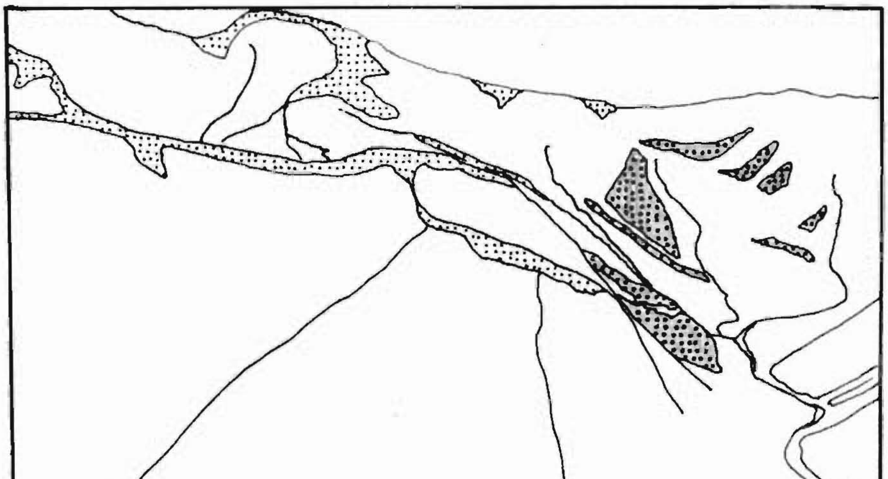


Figure 32c: Hanging Valley Creek showing glacial planed surfaces formed during the early Holocene (upper Tuke Valley).

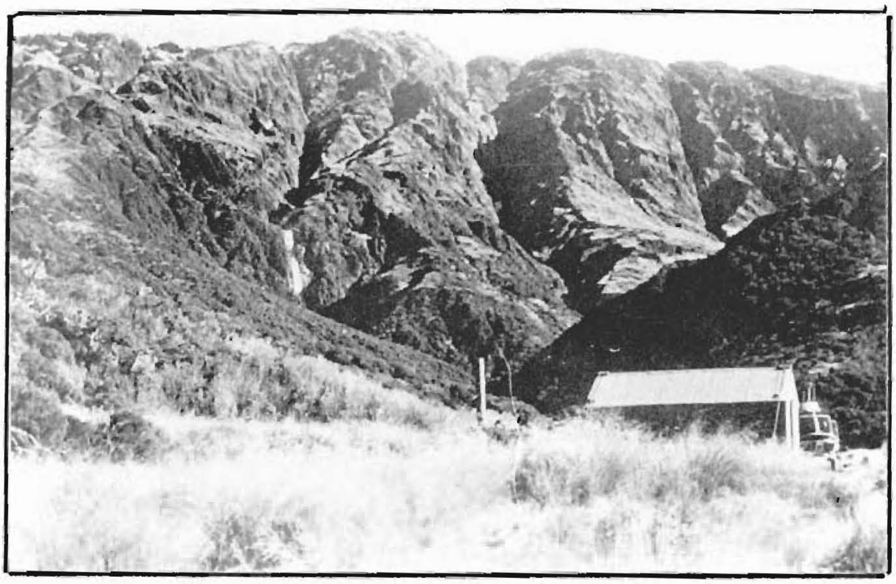
(c) Landslide in Hanging Valley Creek formed by collapse of glacially oversteepened slope cut parallel to schistosity.d. Landsliding may have been initiated by removal of lateral support during ice retreat.

c

d

Figure 33: (a) Upper Cropp Valley showing: high-level glacial planed surface on Galena Ridge, Holocene cirque of the Top Cropp, Waterfall draining cirque in Beaumont Basin and fault controlled drainage of Galena Ridge.
(b) View of Upper Cropp from Sentinal Peak showing high level glacial planed surfaces on ridge tops and late Holocene cirque of Beaumont Basin.

a



b

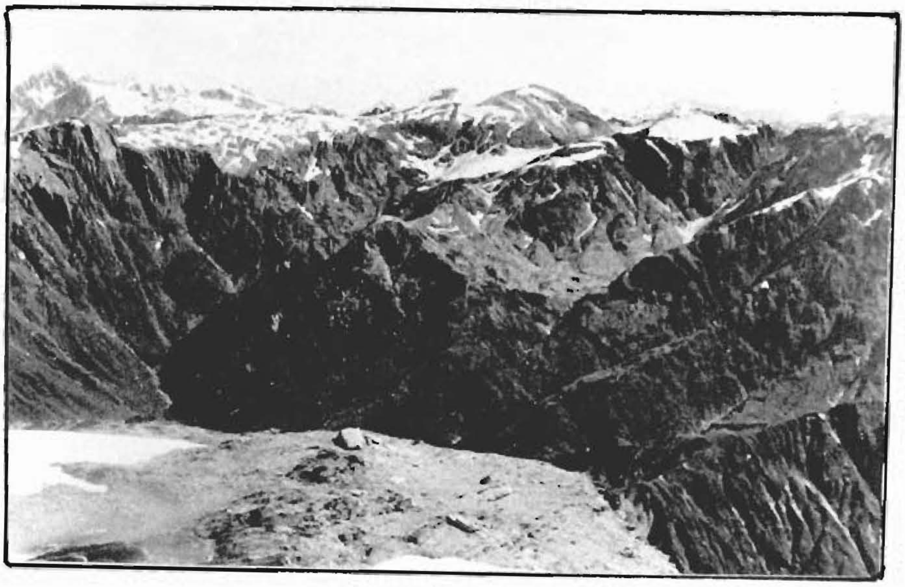


Figure 33(c): High-level glacial planed surface, Remarkable Peak from Galena Ridge.

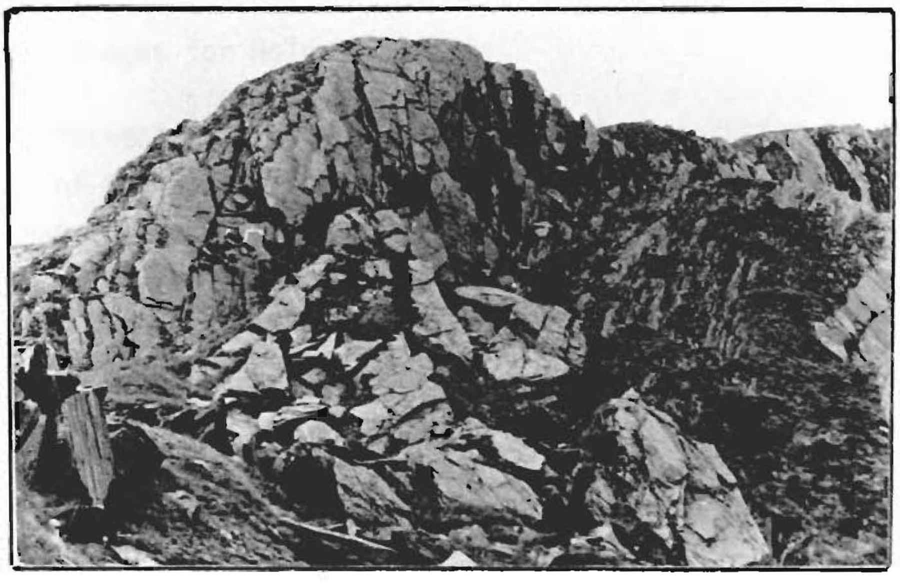
(d) Glacial planed surface Galena Ridge and Mt Beaumont.

(e) Gravity collapse of oversteepened slopes has been one of the major landform modifying processes in high alpine areas since deglaciation.

c

d

e



A major glacial event was associated with the prominent moraine-outwash sequences extending to the sea down the Wanganui and Waitaha Rivers, damming lake Ianthe and pushing outwash gravels down the Totara River. In the Whitcombe-Hokitika area the youngest moraine loop in the Pleistocene sequence is the Moana (Kumara 3, Suggate 1965) terminal at Kaniere Junction. This moraine probably represents the maximum ice cover during the glacial event which produced the high level erosion surface in the Cropp area. The age of the last Pleistocene ice retreat is debatable but a general estimate of 13,500 BP (Chinn 1981) is likely to be correst. Significant modification of slopes in the Cropp area by glaciar ice would have continued for some time after this and an age of 12,000± BP for the glacial surfaces is probably realistic.

Holocene

Cirque basins in Beaumont Basin, Noisey Creek, the Top Cropp basin, Hanging Valley Creek and the top Tuke as well as the ice planed surface of Tarkus Knob represent Holocene glacial events. The incised and narrow gorges of the Tuke and Whitcombe Rivers suggest that major ice movement was restricted to the Cropp and upper Tuke with "dead" ice accumulating in the Whitcombe from icefalls from the Cropp and Vincent Creeks.

The small cirques now prominent in the landscape including the valleys of Vincent and Cataract Creeks were also excavated during this post-glacial advance or series of advances.

Buried peat and soil horizons on surfaces developed during this phase of glaciation indicate that cirque development is not active at present and show promise of giving dates of the end of a major post-glacial cooling event. Until such dates become available only indirect evidence can be used to suggest ages for Holocene events.

Adams (1979) recognised a widespread high level aggradation terrace in the lower levels of the large rivers of Central Westland including the Hokitika. He claims that the terraces were produced synchronously during a phase of aggregation following the glacial advance which formed the Waiho Loop moraine on the Franz Josef Glacier, correlated with the Birch Hill advance apparently dated at 8,500 years. Adams refers to this glacial advance as having extended ice down the Whataroa River past the Alpine Fault to Te Taho.

No evidence was found on aerial photos or 1:63,360 maps for an ice terminus near Te Taho. This is the only valley where Adams infers the advance beyond the alpine valleys combined with the post-glacial aggradation surface. If ice advance did not reach beyond the Whataroa Valley during the Waiho Loop glacial event it is possible that the high level terrace is an outwash surface originating from glaciers entrapped within the alpine valleys. Fig. 34 is an oblique aerial view down the Whataroa Valley showing possible moraines associated with Adam's high level aggradation terrace. It is considered more likely that a widespread aggradation surface would result from glacial advance than from glacial retreat and that the high level terrace referred to above relates to Holocene glacial advance within the Alpine valleys.

If this terrace is a Waiho Loop advance correlative it is dated at 11,000± BP (Chinn 1981). Other possible correlatives are the Birch Hill advance in the Macauley River (in Adams 1979) at 8,500 BP and the McGrath advance in Arthurs Pass (Chinn 1975) at 9,000 BP. Dates from Holocene glacial events in Westland are very rare and an age of 9,000±1,000 years is considered probable in view of the regional chronology and the fact that the high terrace is probably a complex feature relating in part to all three of the above advances.

In the Whitcombe River this terrace building event has probable correlatives in the high level (300 m) fan-terrace complex of Brow Creek (Fig. 35) and below Frew Creek. The wide unterraced valley of the Whitcombe at the Cropp confluence is inferred to have held ice from icefalls out of the Cropp River, Noisey and Vincent Creeks. A high terrace-like feature (on aerial photographs) below Prices Flat suggests that continuous ice did not extend down the Whitcombe during the early Holocene.

Since retreat of the glaciers after 9,000±1000 years there has been no major glacial event producing significant modification of the landscape in either the Tuke or the Cropp Valleys. Glacial activity has been restricted to the cirque basins and the high ridges.

Figure 34: Aerial view down the Whataroa River showing at least two ice front positions (arrowed) associated with the high level outwash terrace. Te Taho where the maximum Holocene ice limit is inferred to have been (J Adams 1979) is behind the ridge on the far right. (Photo from T Chinn, MWD)

Figure 35: Brow Creek crossing the high level Holocene terrace in the Whitcombe River just below the Cropp confluence.

6.4 POST-GLACIAL FEATURES

Post-glacial landform modification has been dominated by gravity collapse of oversteepened slopes and fluvial downcutting through uneven glacial valley profiles.

Gravity collapse of large volumes of bedrock is common in the high alpine area (Fig. 36). This mass movement phenomena occurs when bedrock with a well developed vertical planar fissility which is at a low angle to the trend of a glacially oversteepened valley wall collapses with only minor reorientation of the individual rock slabs. The general effect is one of dilatency associated with removal of lateral support, and such mass movements commonly involve about 10^5 m^3 moving 50 m downslope.

Rapid fluvial downcutting of the main streams and their tributaries has lead to an interesting admixture of fluvial and glacial landforms (Fig. 37). A combination of rockfall from the high slopes (Fig. 36), avalanche and fluvial processes has formed the steep, now vegetated, fans on the south side of the Cropp. The post-glacial fluvial system is discussed in some detail in Chapter 11.

6.5 UPLIFT PATTERN IN THE SOUTHERN ALPS

The recent tectonic history of the Southern Alps, especially the uplift history, is a major control on geomorphic processes in an area like the Cropp River area. With the very high rainfall and the high erosion rate large elevations and steep topography can only be maintained in the long-term if there is a high rate of tectonic uplift. Apart from the geologic evidence for high total late Cenozoic uplifts (see Chapter 5.2) there is no direct evidence on which uplift can be determined from the Cropp area.

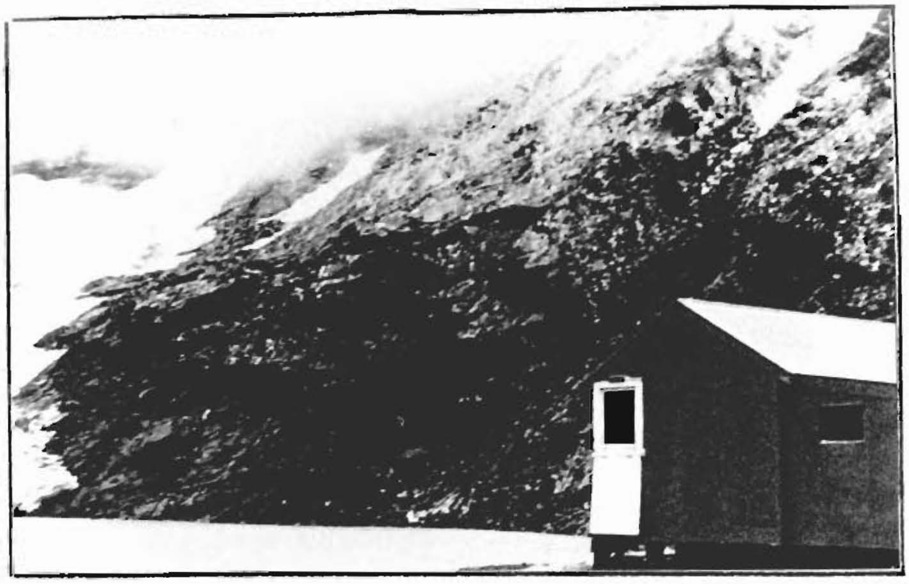
Late Pleistocene valley shape in the Tuke River gives some indication of uplift. The lowest probably glacial slope facet preserved in the Tuke is a slope break in a ridge above the middle gorge, at 700 metres. This is inferred to have been part of the latest Pleistocene glacial valley floor. It is now at least 300 m above the glacial valley floor of the Mikonui River 2 km away, across the Alpine Fault. As the Tuke was the main contributor of ice to the Mikonui Glacier an elevation difference of 300 m is more than would be expected. If half of the height difference is original and

Figure 36: Mass movement in alpine areas is dominated by gravity collapse.

(a) At Ivory Glacier large-scale gravity collapse has followed removal of lateral support during ice retreat.

(b) Faces cut parallel to schistosity are prone to rock fall. Remarkable Peak.

(c) On ridge tops rocks with initially vertical schistosity become "tired" and rest in reclined positions.

a


b

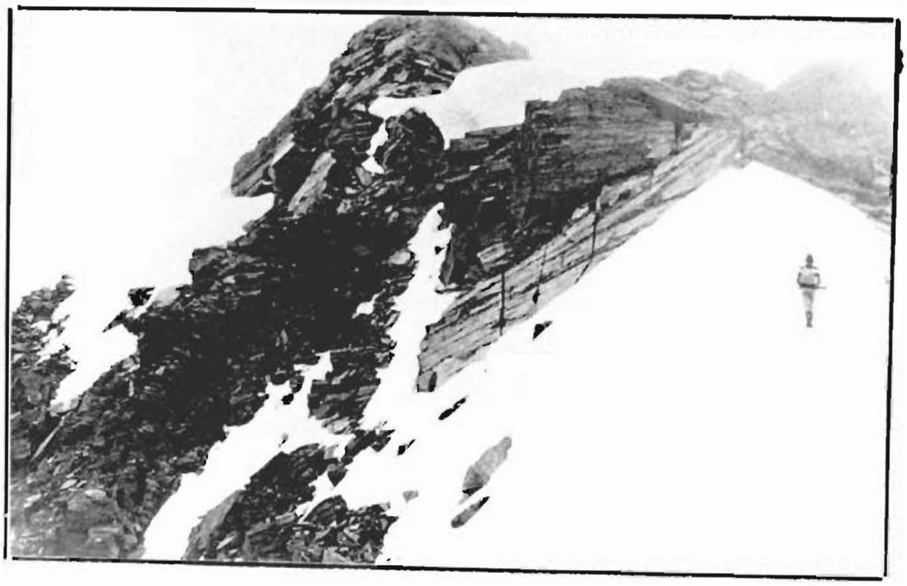
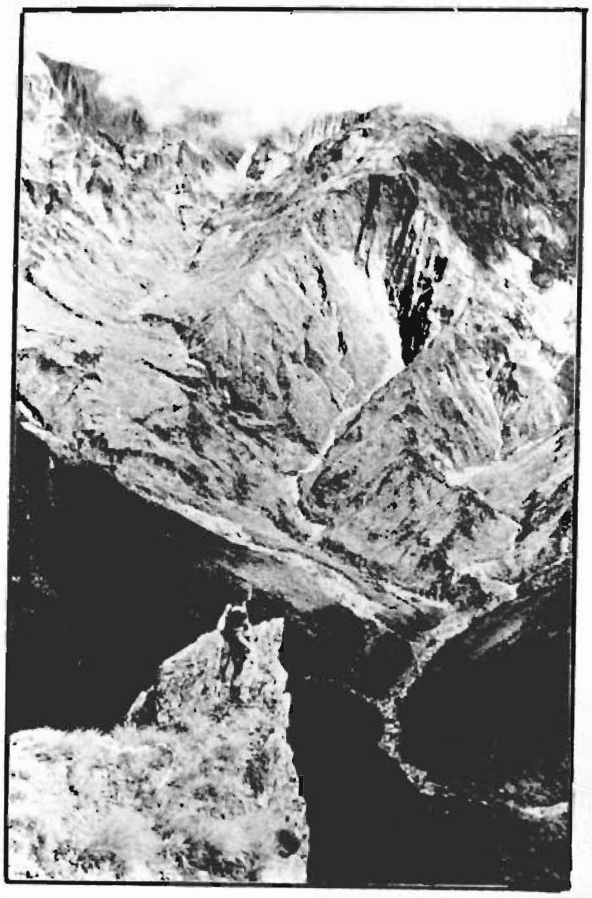

c


Figure 37: The geomorphology of the Cropp combines elements of glacial (cirque at Top Cropp (a)) and entirely fluvial (Limpopo River (b)) landforms.

a



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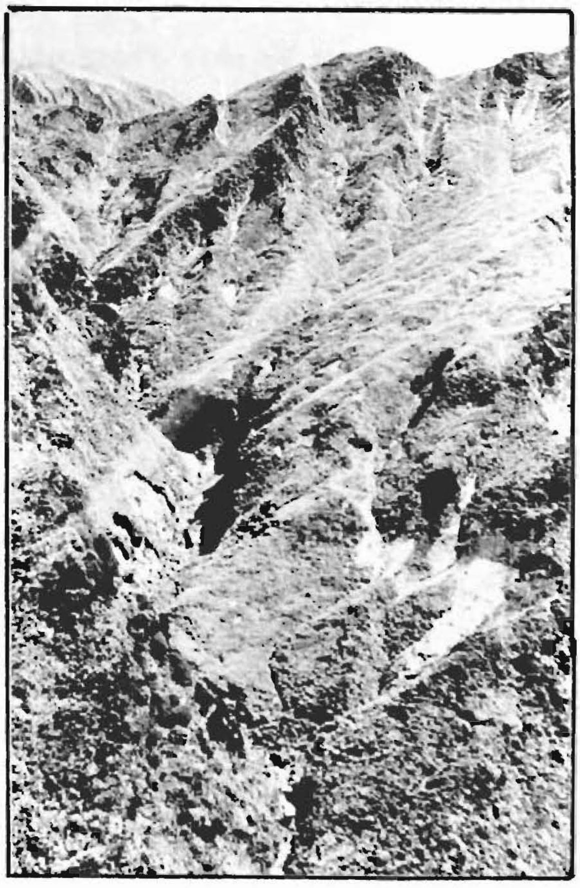


half tectonic the uplift rate would be 12.5 mm per year over the last 12,000± years.

All along the Alpine Fault direct evidence for fault movement has been observed. Apart from minor anticlinal warps of unconsolidated sediments overlying the Alpine Fault zone (Suggate 1968) uplift in the Southern Alps is considered by the writer to be a block uplift localised along faults in the east and the Alpine Fault in the west. The suggestion that uplift distribution is affected by vertical drag on the Alpine Fault and that uplift is maximised 5 km southeast of the fault (Adams 1979) is unproved and is based on equivocal interpretations of sparse geomorphic data.

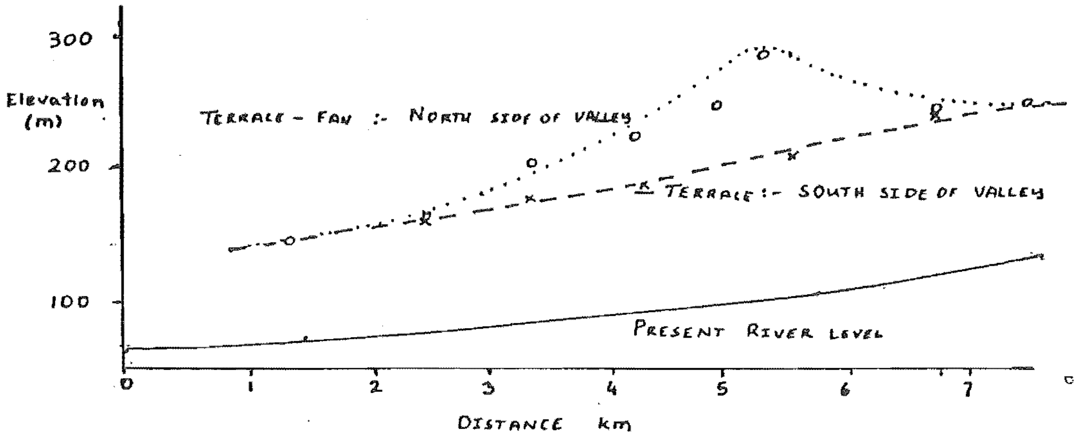
Adams (1979) cites two lines of evidence for vertical drag on the Alpine Fault and concentration of uplift southeast of the fault. There is a high level terrace present in The Wanganui, Whataroa, Paringa and Okura Rivers (see section 6.3) which is considered by Adams (1979) to have been deformed into anticlinal warps. In the Wanganui River the terrace level figures by Adams as being a deformed originally planar surface is more likely part of a fan-terrace complex with high original dips on the surface. There is an undeformed terrace level traceable on the south side of the Wanganui River which maintains a constant gradient from the Alpine Fault to 10 km upstream (Fig. 38). From 1:63,360 maps and aerial photographs the terrace sequence in the Whataroa River appears to be obscured by fan and landslide surfaces (of the same generation as the terrace level) but appears to be undeformed. The Paringa anticlinal structure is described by Suggate (1968) but evidence for a deformed agradation surface is not well documented.

Displacement across "schistosity plane faults" is cited in Adams (1979) from the Franz Josef area as the mechanism for producing the inferred vertical drag features. Displacements as described require no horizontal movement (Fig. 38), are inconsistent with a tectonic regime involving surface faulting at oblique angles to schistosity (Chapter 4.2 and Findlay 1980) and are well within the scale of displacements which might be expected from isostatic rebound following removal of 70 m+ of ice. The concept of vertical drag on the Alpine Fault is unproven and uplift rates inferred from such a model (20±4 mm per year at Wanganui River, Adams 1979) will not be considered in further discussion of uplift of the Southern Alps.

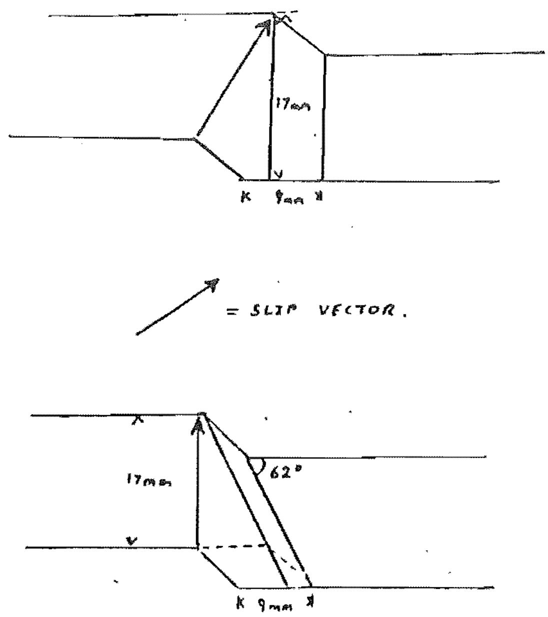
Fig 38:Alternative interpretations of evidence suggesting differential uplift across the schist belt.A:J Adams 1979 uses a supposedly deformed terrace in the Wanganui River to suggest uplift is localised 5km East of the Alpine Fault. Alternatively this surface could be a terrace-fan complex.From 1:63360 maps and aerial photos an undeformed fluvial terrace may be present on the South side of the Wanganui valley,of similar age to the fan-terrace complex on the North.

B.Apparent horizontal offset of schistosity plane faults (Adams1979)could be caused by a dipping fracture plane associated with vertical isostatic rebound.

A EARLY HOLOCENE TERRACES : WANGANUI RIVER.



B



The high level Holocene agradation surface described by Adams (1979) has probable correlatives in the Toaroha, Hokitika, Wanganui, Whataroa and Karangarua Rivers with possible correlatives in the Mahitahi, Paringa, Moeraki, and Haast Rivers from air photo and map interpretation. From the Hokitika southward the height of these terraces above the present river floodplains decreases regularly from 100 m+ at the Hokitika till it merges with the present floodplain south of the Karangarua River. If these terraces are eventually synchronous, as is suggested by their association with Holocene moraines in the Whataroa and Karangarua Rivers, and the elevation across the Alpine Fault is tectonic then the constant height variation (Fig. 39) suggests that uplift across the Alpine Fault has been greater in Central Westland than in South Westland. The alternative, that the surfaces are not synchronous or that elevation is not tectonic is difficult to reconcile with the regular elevation change.

To check the hypothesis that vertical throw on the Alpine Fault increases northward calculated uplift rates (from Wellman 1979, Fig. 40) across the Alpine Fault, excluding the rates derived using a "vertical drag" model, were plotted against distance along the fault. The data are surprisingly consistent indicating a decrease in uplift of 3 mm per year for every 100 km south of Hokitika.

Such an uplift pattern is entirely consistent with the observation of late Tertiary schists in the Cropp area and the lack of young schists to the south (see Chapter 5.2). This is also consistent with plate tectonic predictions that the direction of plate convergence is more oblique and longer-lived in Central than in South Westland (Walcott 1978). North of the Taramakau it is assumed that compression is relieved by movements on the Alpine Fault branch fault system with significant rotation (Walcott <u>et al.</u> 1981) and uplift is correspondingly reduced.

One implication of the inferred uplift pattern is that the area of maximum uplift in the Southern Alps is probably not in the Mt Cook area where elevations are highest, but in the Hokitika region where high denudation rates have kept elevations relatively low.

Fig 39:High level river terraces offset at the Alpine Fault show a decreasing apparent offset with distance south.Two of these terraces are associated with Holocene moraines,suggesting that they are outwash surfaces.

PROFILES OF RIVER TERRACES -WESTLAND

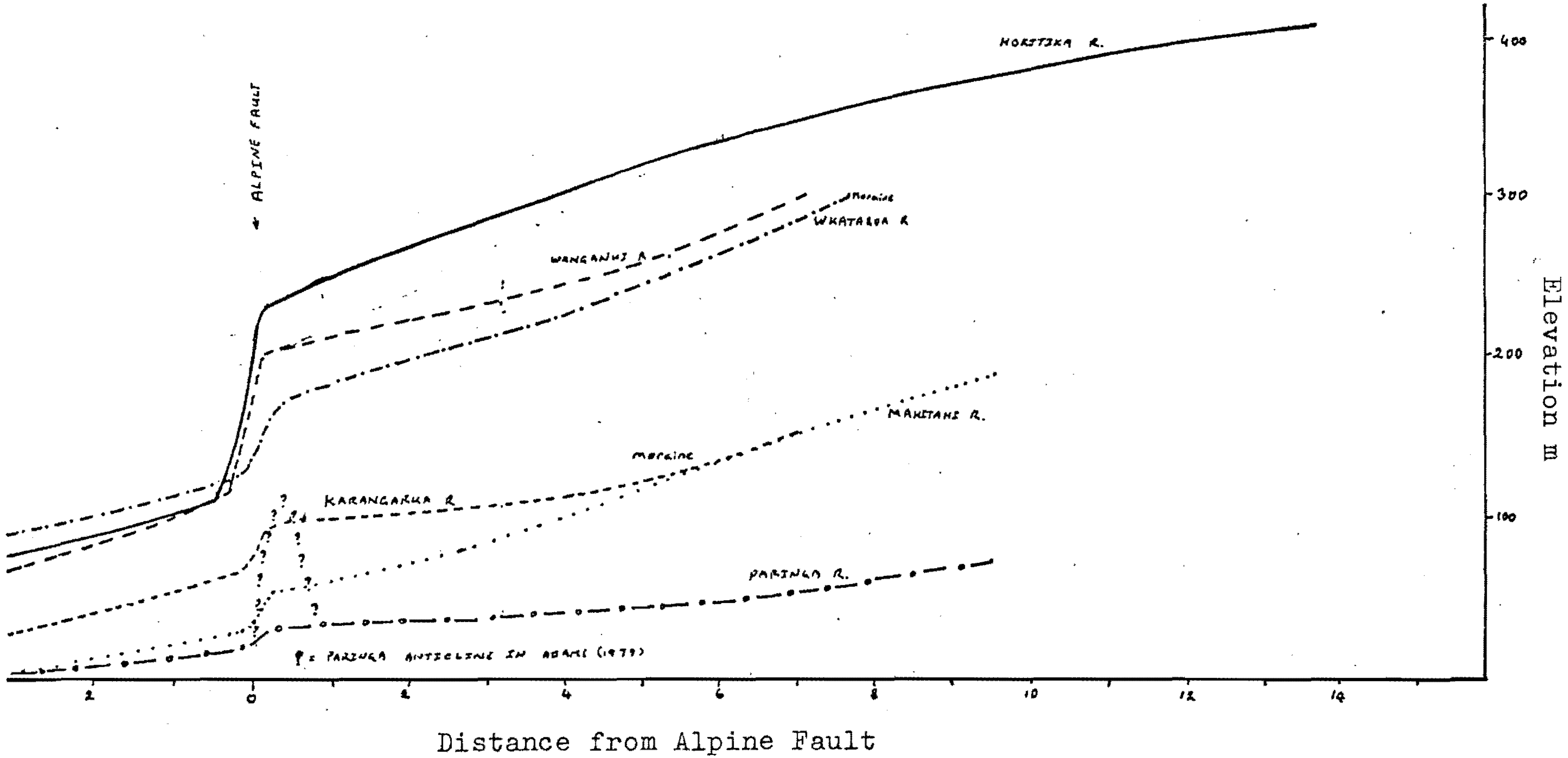
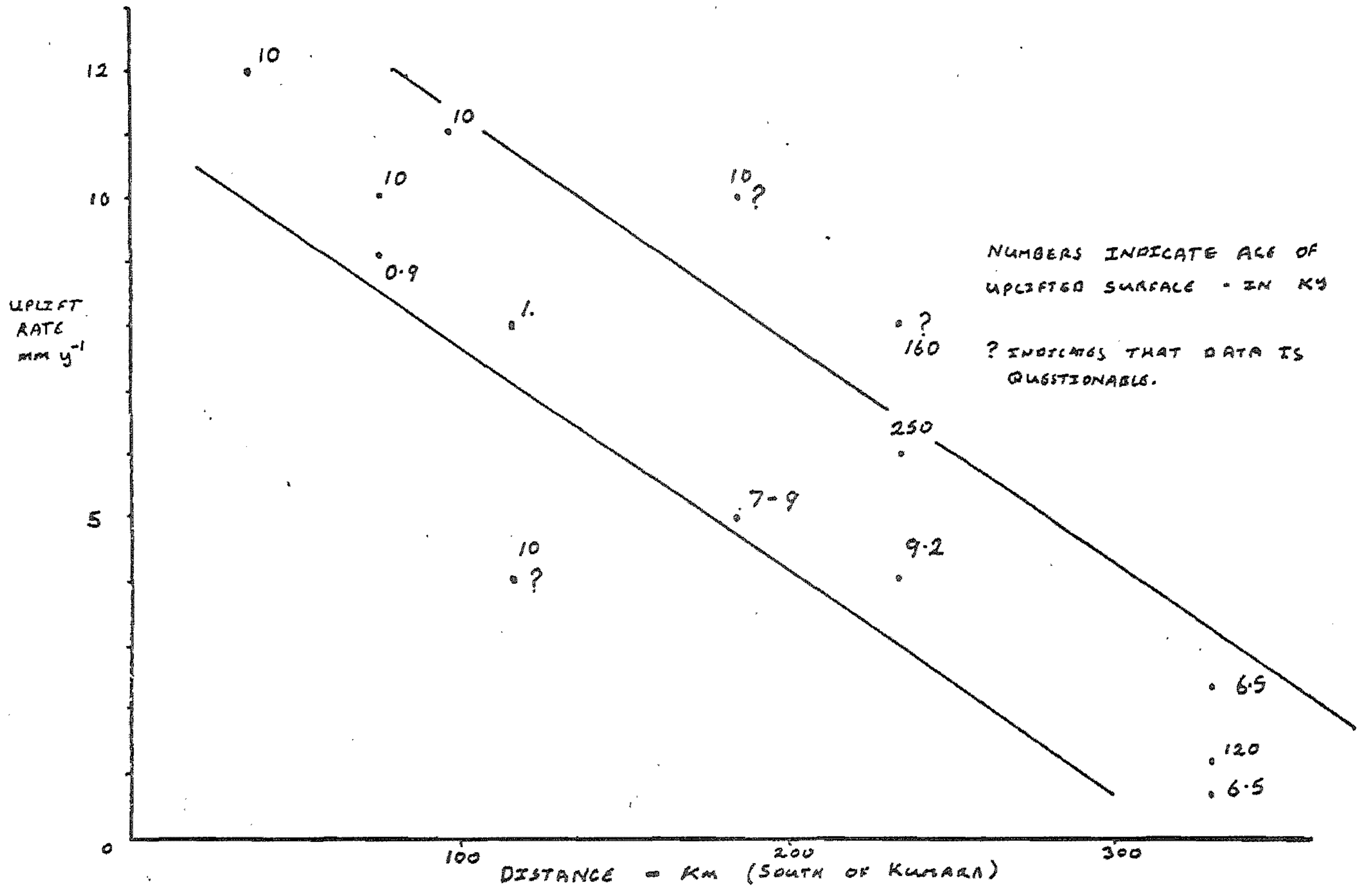


Fig 40:Uplift rates on south-east of the Alpine Fault.
Values from Wellman 1979.



Geologic and geomorphic considerations imply total late Cenozoic uplift for the Cropp River area of 12 to 24 km (12 km from K/Ar dating, 24 km from 2 Ma of uplift at 12 mm per year). Uplift rate has been increasing with time as the pole of rotation for the Pacific plate relative to the Indian plate has migrated southward in the late Cenozoic (Walcott 1979). The present uplift near the Hokitika River is probably near the late Cenozoic maximum regional uplift rate for the Southern Alps 12 ± 4 mm yr^{-1}.

Differential uplift is probably accompanied by discrete fault plane movement near the surface and by synmetamorphic folding at relatively shallow depths. Because of the high geothermal gradient only the shallow faults need be seismic and stable creep conditions may occur at shallow depths.

"As a general principle it is probably best not to get involved in sediment yield (measurement) programmes."

A.J. Sutherland 1978

PART II

SEDIMENT YIELD OF THE CROPP RIVER

CHAPTER SEVEN: SYNOPSIS

The following is a study of an extreme environment. Results of this study indicate rainfall, runoff, erosion rates, denudation rates and uplift rates comparable with the highest measured anywhere in the world. However, it is also a study of a catchment which has adjusted itself to cope with both short and long-term extremes. Hopefully this study will provide an endpoint against which less extreme environments can be compared.

Over a one-year period observations were made on rainfall, hydrology (water yield), hillslope erosion processes, suspended sediment yield and features of bedload movement, combined with a geomorphological study, in the upper Cropp River catchment. The main aim of this work was to assess erosion processes and rates as reflected by sediment yield and especially suspended sediment yields in the Cropp area which is distinguished by having a very high annual rainfall (10 m+) and a high geological uplift rate (12$\pm$ mm yr^{-1}).

Climatic observations were restricted to using the rainfall data collected by the Ministry of Works and Development for the Cropp and surrounding areas and comparing this data with the long-term record for Hokitika Airport. Rainfall measurements showed a rainfall variation within the area from 8.6 to 10.5 m with an average of about 10 metres. The Hokitika record for 1980 gave a rainfall of 2672 mm which is 4.6% less than the annual average (1964-1979). The 1979-1980 study period involved 43 days with rainfall in excess of 100 mm at the Cropp Hut. Only one of these storm events is likely to have been unusual. This was because of its long duration rather than abnormal intensity. As far as could be determined rainfall conditions during the study period were reasonably representative of the long-term rainfall in the Cropp.

The hydrology of the 12.2 km^2 upper Cropp River catchment was studied using data collected at the middle gorge of the Cropp and data from six sub-catchments of the upper Cropp. A stage height-discharge rating curve derived for the Cropp at gorge section accurately predicted water yields comparable to rainfall for individual storm events but overpredicted annual runoff because of sensitivity of the relation to minor bed form changes at low flows.

The most striking feature of upper Cropp hydrology is the very rapid hydrograph response. In an extreme example discharge of the Cropp rose from twice mean flow to 100 times mean flow in three hours and returned to twice mean flow within four hours.

Unfortunately during the period when most observations were made (January/ February 1980) the gauging site on the Hokitika River had a major siltation problem. As a result only general correlations could be made between Cropp River and Hokitika River flow conditions.

Suspended sediment samples collected over a range of flow conditions (up to 100 times mean flow) allowed derivation of a discharge-sediment concentration rating curve. Comparison with discharge records gave an estimated annual suspended sediment yield for the upper Cropp of 32,400$\pm$11,290 t/km^2 per year for 1980, representing a denudation rate of 12.2$\pm$4.31 mm per year.

A study of slope conditions and erosion processes in a small (4.5 ha) area was undertaken to relate hillslope erosion patterns and trends and stream sediment yields. The catchment was selected because of accessibility and size and is considered to be a typical upper Cropp grasslands catchment. This study showed that most erosion features supplied sediment directly to the stream system and that material eroded from bedrock exposures on slopes could be, and was, transported into the major tributaries within one storm event. Although evidence of rapid erosion was observed on all erosion assessment sites no removal of material was systematically measured largely because of inadequate measurement techniques. It would seem that once initiated, erosion features on steep slopes in the Cropp area rapidly attain an equilibrium configuration which allows them to continue to supply sediment to the river system probably for tens of years. Stabilisation usually occurs as a result of changes in the stream rather than because of on-slope processes.

In conjunction with study of sediment sources and suspended sediment yields bedload movement in the upper Cropp was also examined. Because of the impracticability of direct measurements study of bedload yield of a small catchment was combined with estimation of relative bedload contribution of tributary streams based on gravel composition and experimental abrasion of bedload samples in order to estimate the amount and relative importance of bedload to the total sediment load. This aim was not fulfilled because bedload movement in the Cropp does not appear to conform to a simple transport-abrasion model. Observations both from experimental studies and observations of flood events suggest that bedload movement is significant at high discharges but that abrasion, and more significantly grinding and impact breakage of pebbles can convert the highly fissile schist gravels to a largely suspended load in a very short transport distance.

The results of the study of present-day processes were then compared with late Holocene erosion history as it is reflected in the geomorphology of the Cropp basin. Identification of a latest Pleistocene glacier cut erosion surface on ridge tops throughout the Cropp Valley allowed the recreation of a series of 12,000 BP cross profiles of the valley. When combined with the present-day cross profiles the volume of material removed between 12,000$\pm$ BP and the present can be calculated for the Cropp and its sub-catchments. The overall denudation rate implied by this study is 13.3$\pm$3.4 mm per year representing removal of 35,200$\pm$9,010 t/km^2 per year from the upper Cropp catchment.

Sub-catchments of the Cropp show varying preservation of relict glacial topography depending on the amount of post-glacial fluvial dissection and downcutting. The areas with greatest fluvial downcutting show the greatest long-term denudation effects and those areas where there is some physical barrier to river entrenchment, denudation rates are suppressed. This suggests that the greatest control on denudation is the rate of downcutting of streams controlling the base level of hillslopes rather than hillslope processes.

Studies of sediment yield in the Cropp River area indicate that present-day catchment conditions are essentially the same as average conditions for the last 12,000 years, with respect to sediment yield. Sediment yield is of the order of 35,200$\pm$9,010 t/km^2 per year over the long-term and 32,400$\pm$5,747 t/km^2 per year over the one-year study. Differences between these figures represent the unmeasured bedload component of the short-term sediment yield which need not be greater than 10% of total load from these figures. The majority of sediment removed from the Cropp travels in suspension. Thus suspended sediment records give a reliable estimate of denudation within the catchment. On the scale of the sub-catchments studied in the Cropp sediment yield is controlled by geology (lithology, structure and tectonics) and basin configuation and shows no measurable effect of climatic variation across the catchment.

Comparison of the nature of sediment produced by the Cropp system (coarse sand to silt-sized schist fragments in suspension and relatively minor amounts of highly fissile quartzofeldspathic schist gravel) with sediments in the lower Hokitika river system (coarse, greywacke dominated gravels with very minor schist and with sand and silt only deposited in ponding areas during waning flow conditions) suggests:

a) that most of the sediment derived from the Cropp is not deposited within the Hokitika system but is passed out to sea, probably within one single storm event, as suspended sediment (a travel distance of 60 km within flood events typically of less than 48 hours duration); and

b) that Cropp derived sediments are not a cause of aggradation or river control problems in the Hokitika River.

CHAPTER EIGHT: CLIMATE

The Cropp River lies within a narrow zone of extreme alpine climate on the western front ranges of the Southern Alps. This region is characterised by high and highly variable rainfall with annual totals locally exceeding 12 metres.

The basic control on rainfall seems to be the distance from the westernmost 1500 m topographic contour (McSaveney pers. comm. 1980). The rainfall maximum is in a belt parallel to the topographic maximum but 5 to 10 km to the west, across the head of the Cropp River in this area.

Rainfall records are kept by the MWD for five sites within the Cropp and for various stations in nearby catchments including a continuous recording tipping bucket gauge at Cropp Hut. Monthly rainfall variation is up to 50% between stations in the Cropp with the hut site consistently drier than the upper catchment gauges. A map of rainfall distribution across the area for 1977-1979 is shown in Fig. 41.

Main features of the 1980-1981 rainfall measured at Cropp Hut included 205 rain days, with 43 days between September 1979 and October 1980 receiving in excess of 100 mm per day and with a daily maximum of 464 mm reached on 24 December 1979. Intensities greater than 25 mm per hour occurred regularly with a recorded maximum of 49.5 mm per hour (see Table 5).

Daily rainfall at Cropp Hut is generally five to ten times that recorded at Hokitika Airport, the nearest station with reliable long-term records. 1980 rainfall at Hokitika was 2672 mm, 4.6% less than the 1964 to 1979 mean of 2791 mm. It is assumed that Cropp annual rainfall for the 1980-1981 year is also within 10% of the long-term mean. Flood return frequency analysis on the Hokitika River showed that one flood during the study period produced a discharge with return frequencies exceeding one year (see Chapter 9). The storm involved contributed less than 3% of the total precipitation at Cropp Hut for the year.

Meteorologically the 1980-1981 period was not unusual for the Cropp area from comparison of regional rainfall and river flow records.

Allowance must be made in the Cropp rainfall data for catch losses of the rain gauges (believed to be less than 5%) for snowfall where catch efficiency of the rain gauges is unknown and for periods when rain gauges have overflowed. The effects of long-term snow storage of precipitation and of fog precipitation are assumed to be minimal. Given that all these inaccuracies minimise the measured value an acceptable precipitation value for the upper

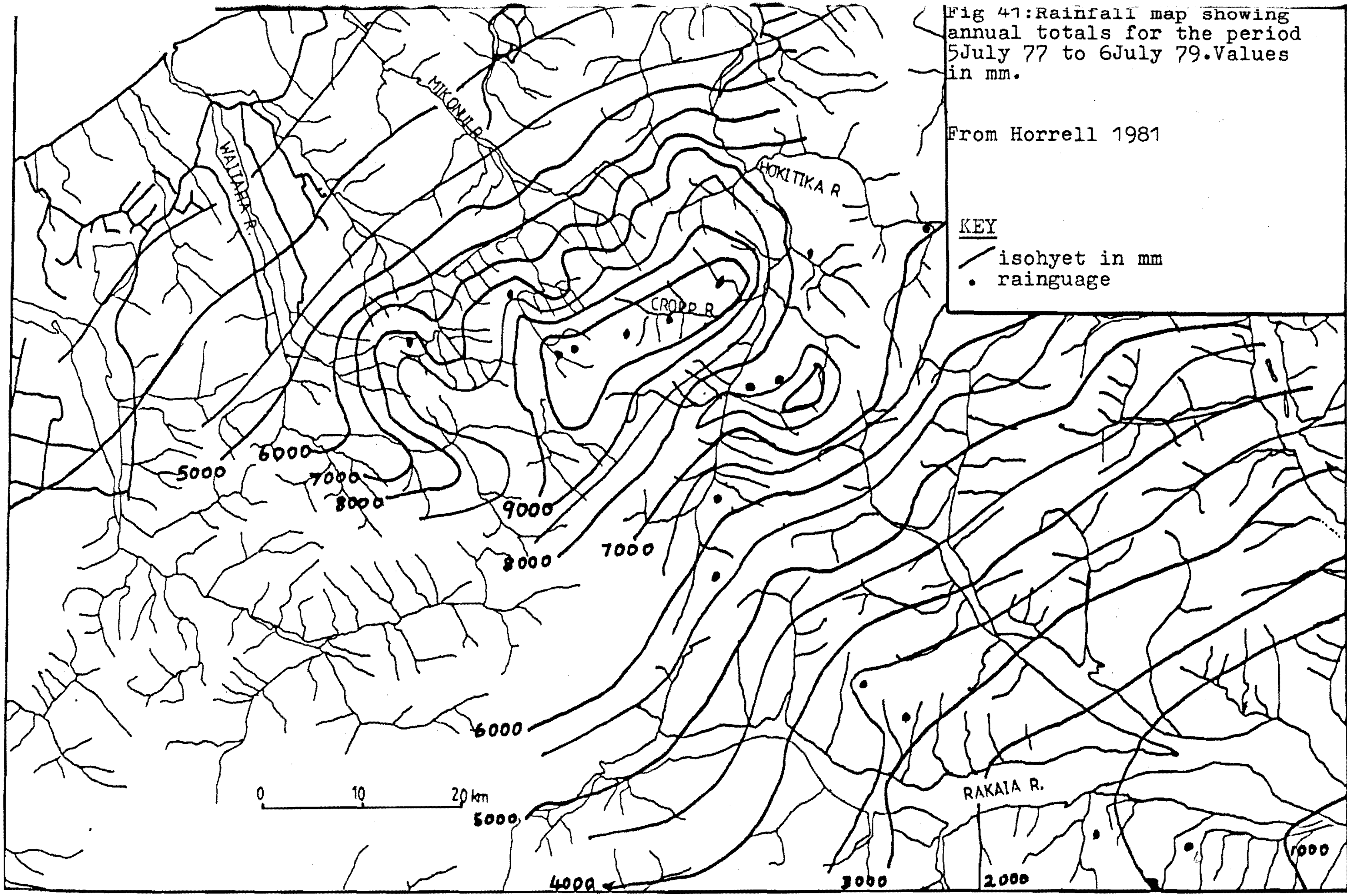


Fig 41:Rainfall map showing annual totals for the period 5July 77 to 6July 79.Values in mm.

From Horrell 1981

KEY

isohyet in mm

• rainguage

TABLE 5: Raindays exceeding 100mm rainfall 29/9/79 to 26/8/80 at Cropp hut compared with rainfall at Tuke hut and Hokitika Airport. All figures in mm.

HOKITIKA RAINFALL	TUKE RAINFALL	CROPP RAINFALL	
9·3	-	136	29/9/79
30·6	-	175	30/9/79
9	-	129	12/10/79
29·7	-	194	16/10/79
56·4	-	144	28/10/79
57·6	-	247	29/10/79
1·2	-	107	30/10/79
33	80	114	17/11/79
13·2	174	108	22/11/79
61·7	48	133·5	27/11/79
36·8	228·5	118	28/11/79
3·5	88·5	100	30/11/79
1·7	47·5	113	1/12/79
54·1	477	380	2/12/79
20	143	151	18/12/79
8·6	100	103	20/12/79
30·9	93·5	108	22/12/79
10·7	538	464	24/12/79
-	181	199	25/12/79
34	143	134	15/ 1/80
27	183	200	16/ 1/80
25·6	134	143	19/ 1/80
18·7	107	178	24/ 1/80
9·6	250	221	27/ 1/80
18·3	14	143	3/ 2/80
2·9	111	144	4/ 2/80
13·5	149	163	25/ 2/80
29·8	94	132	15/ 3/80
14·3	85	154	16/ 3/80
6·2	84	149	20/ 3/80
19·6	121·5	166	30/ 3/80
71·5	120	166	31/ 3/80
39	93	114	18/ 5/80
23·5	95	104	20/5 /80
15·1	44	106	21/ 5/80
15·1	164	110	22/ 5/80
77·2	119+	327	10/ 6/80
8·7	-	110	30/ 6/80
12·9	-	170	12/ 8/80
12·4	-	115	13/ 8/80
39·5	-	217	15/ 8/80
2·8	-	140	16/ 8/80
48·8	-	183	26/ 8/80

RAINFALL INTENSITY

10hr	5hr	1hr	30min	12min
102	55	18·5	10	5·5
139·5	70	25	15	9·5
138	105	49·5	30·5	17·5
170	136	46·5	33	20·5
221	163	47	24	11·5

Cropp for 1980-1981 is considered to be 11 ± 1 m with 10 m as a minimum possible value.

Wind is another feature of the climate of the Cropp. Although no direct measurements of wind were taken some general observations on the effects of wind especially in the open tops deserve mention, as the erosive power of strong winds is often neglected because of measurement difficulties.

In the storm of 4 December 1979 the New Zealand Forest Service's shelter "Noisey biv" was blown off its foundations and about 50 m downslope. The hut, plus contents weighed an estimated 200 kg, was in a relatively sheltered position and suffered no structural damage during its journey. Other huts above the bush line have met a similar fate (ie Cat Creek in the Whitcombe Valley). On the divide at the head of Noisey Creek is a steep face of disrupted bedrock dropping into Rapid Creek and facing northwest. On the Noisey Creek side of the divide is an easy sloping area of snowgrass containing no obvious sediment sources or catchment areas for producing significant overland flow. Small piles of schist pebbles occur within the vegetation mat and on the southeast (leeward) side of obstacles. Blades of schist up to 10 cm (long dimension) have been observed lying on live vegetation. Casual observation lead to the conclusion that these sediments were wind deposited having been blown off the steep face at the head of Rapid Creek, over the divide and deposited in snowgrass "baffles", a transport distance of several metres.

The presence of strong winds does not directly affect this study except in as much as conventional rain gauge data must be interpreted in view of the complex precipitation patterns likely under gale conditions and that erosion patterns produced by such rainfall may be significantly different from those produced by steady, near vertical rainfall.

CHAPTER NINE: UPPER CROPP HYDROLOGY

The hydrology of small New Zealand alpine catchments regularly subjected to extreme total rainfalls and intensities is very poorly known. Intermittent visual observations of storm events and river responses are equally rare as even the most dedicated observer is more comfortable in a hut or tent when rainfalls exceed 1 mm per minute. Because of time limitations imposed by thesis requirements this study falls somewhere in between hydrological measurement and general observation.

9.1 THE CATCHMENT

The Cropp is a 28.5 km^2 catchment which drains eastward into the Whitcombe River, a major tributary of the Hokitika River (a NWASCO representative basin, see Fig. 42). The upper 12.17 km^2 of this catchment is gauged at the head of a small gorge (S64/524117) and it is this upper catchment which is discussed in the following section.

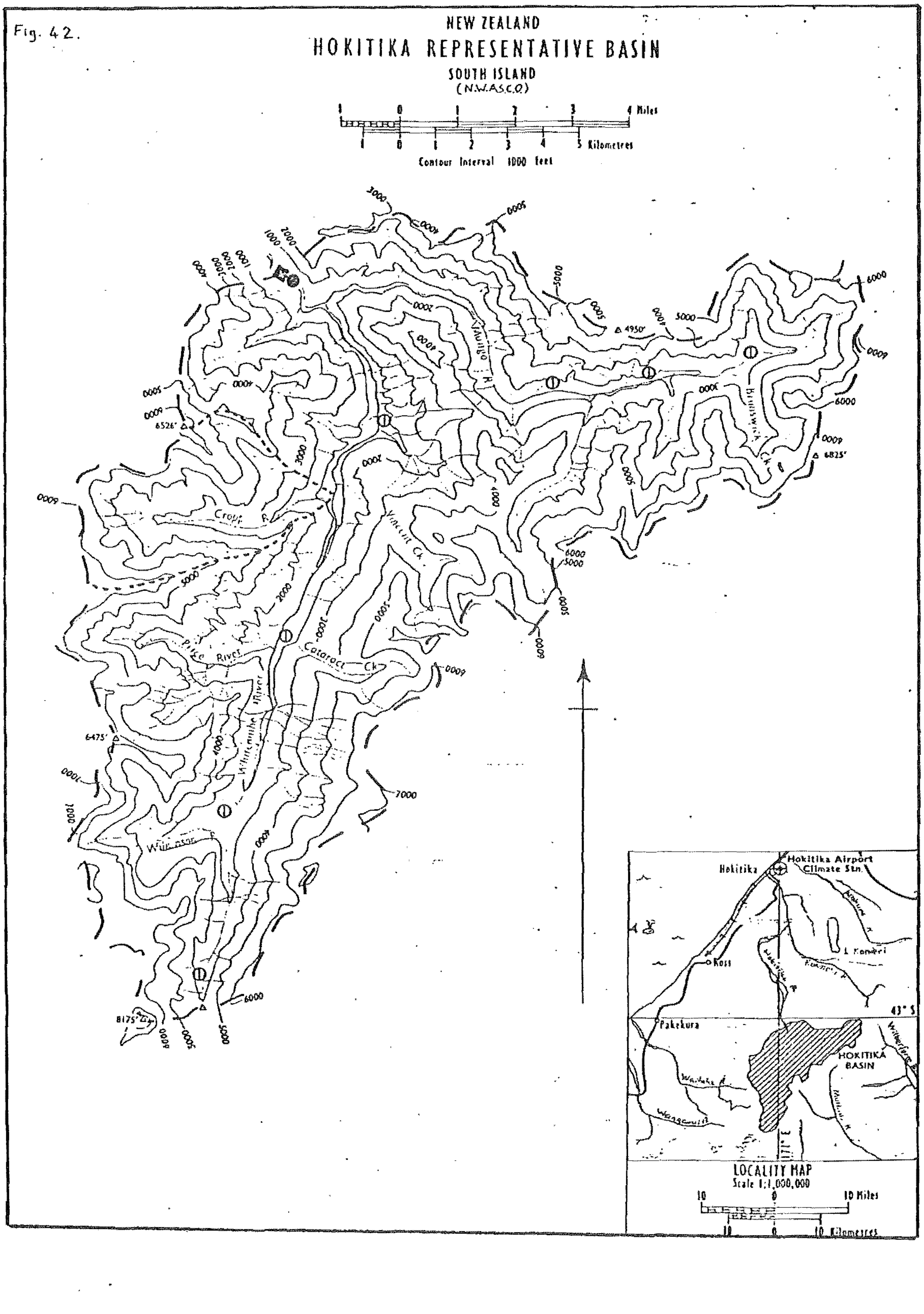
The Cropp River at the gauging site drains a nearly rectangular basin about 3.5 km long and 3.8 km wide. Average slope in the catchment is 29^0 (from average contour spacing on 1:10,000 map), although slopes exceeding 50^0 are common over large areas. Elevation range is 2140 m to 865 metres.

Six sub-catchments were considered in the hydrological and sediment yield studies (see Fig. 43 for catchment locations):

1 The Top Cropp in the northwest corner of the basin;
2 Rotunda Creek, the other major headwater tributary;
3 Mid-Cropp gauged below the confluence of Top Cropp and Rotunda;
4 Whinging Creek draining the steep south side of the Cropp Valley;
5 Danger Gully draining the highly shattered pelitic schist of the north side of the Cropp; and
6 A small catchment within Danger Gully, the Limpopo River.

Various mathematical relations describing the basin shape, size, topography and drainage patterns (Gregory and Walling 1973, Chapter 2) were considered but none gave any fundamental insight into the character of the basin other than that which can be seen on a scaled contour map (Fig. 43b). Drainage density (km of stream length/basin area) was considered but due to the difficulty of definition of what constitutes a stream does not contribute

Fig. 42.


NEW ZEALAND
HOKITIKA REPRESENTATIVE BASIN
SOUTH ISLAND
(N.W.A.S.C.O.)
Miles
Kilometres
Contour Interval 1000 feet
Cropp R.
Vincent Ck.
Price River
Cataract Ck.
Whitcombe River
Wilkinson R.
Brunswick Ck.
LOCALITY MAP
Scale 1:1,000,000
Hokitika
Hokitika Airport Climate Stn.
Ross
Pukekura
HOKITIKA BASIN
43° S
10 Miles
10 Kilometres

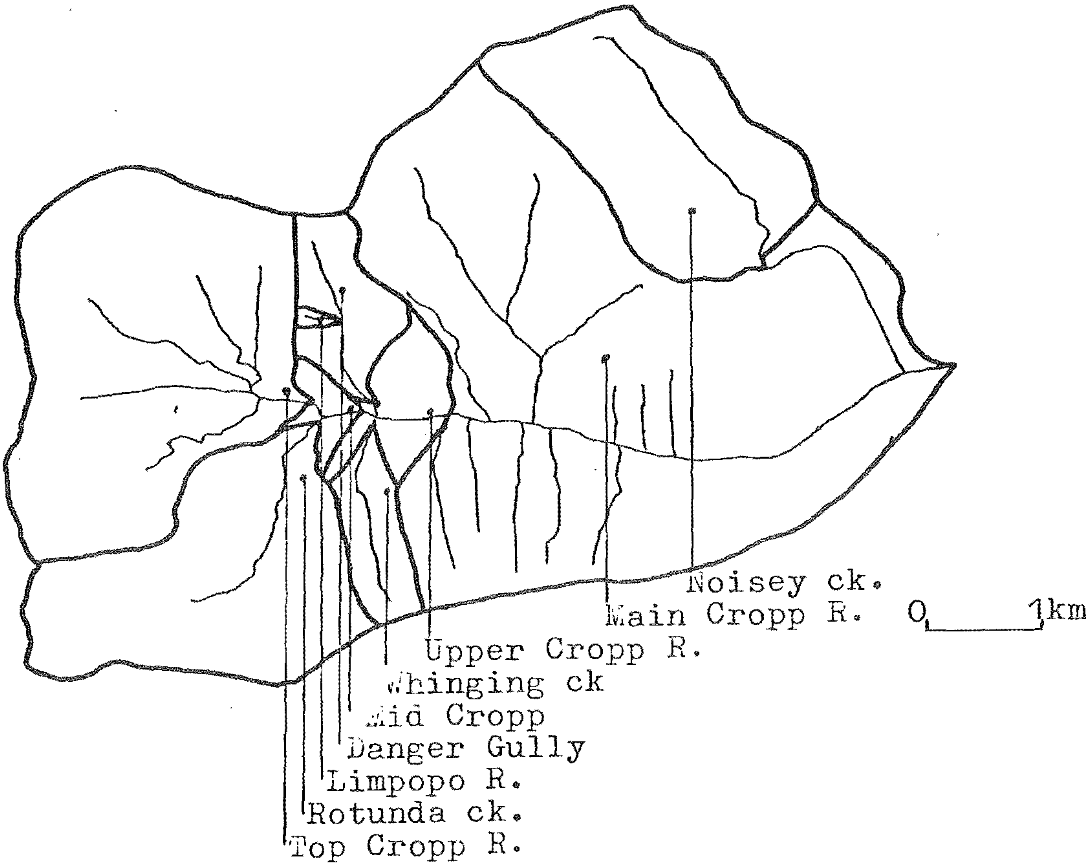


Fig.43a.Sub-catchments of the main Cropp River.

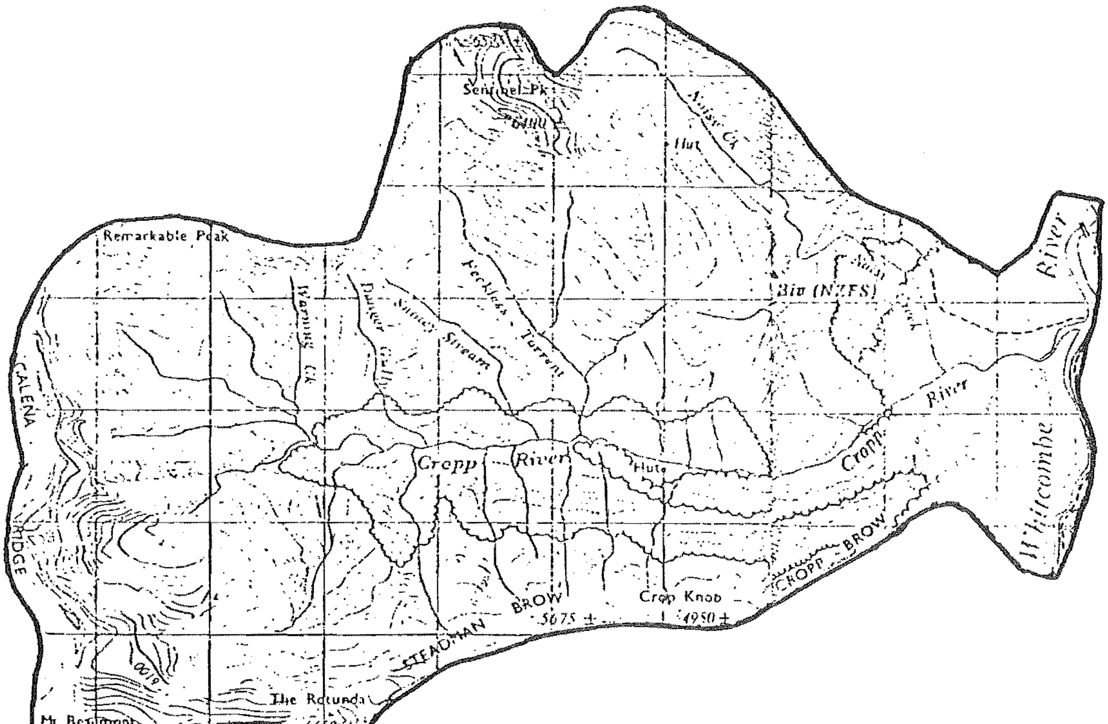


Fig.43b:Cropp River,1:63360,elevations
in feet.From N.Z.M.S.1,S64.

to an understanding of the Cropp system. As most watercourses in the Cropp are ephemeral due to steep slopes and a lack of storage area in most catchments, drainage densities from 2.5 to 5 were determined (from 1:10,000 aerial photographs) depending on whether only perennial streams or all major drainage channels were considered. There is an extensive network of small surface drainage features down to the scale of depressions eroded between individual tussocks on most slopes in the Cropp. If these are considered as significant drainage features, which they are during storm events, the drainage density would increase by at least an order of magnitude. The very rapid hydrograph response of the Cropp River suggests a high drainage density value.

9.2 STREAM FLOW MEASUREMENT

Because the majority of sediment removed from rivers such as the Cropp is transported as suspended sediment a record of stream discharge was necessary to allow calculation of suspended sediment transport rates and yield. Suspended sediment yield of sub-catchments within the Cropp is also assumed to be proportional to catchment water yield. Features of flood hydrographs are also of interest as they indicate possible processes active in the catchment during storm events.

Stream discharge estimates have been made for the Cropp River at the gorge site and for each of the sub-catchments using wading rod gauging for low flows and maximum surface velocity in surveyed sections for medium and high flows (Toebus 1963). Suspended sediments were sampled for high flows on all sub-catchments but because of access difficulties full stage/discharge and stage/sediment concentration curves were only produced for the "Cropp at gorge" section.

Stage height records for the Cropp Gorge are kept by the MWD from a stilling well and five minute digital punchout water level recorders (Fig. 44). Flow gaugings at low flows (flows up to twice the mean flow) were done using a Gurley vertical axis current meter and wading rod (Fig. 45) and at high flows using surface floats to measure maximum surface velocity. As a further check a rating curve was produced using the Manning equation (Barnes 1967). The Manning equation was found to underestimate stream flows considerably at high flows because of uncertainties in the water surface slope and the nature of the bed during high flows when sediment movement is significant (see Simon et al. 1979 and Chapter 10.3).

Figure 44: Gauging tower at Cropp, at Gorge site during:

(a) flood of approximately one month return frequency;

(b) flood of about one year return frequency;

(c) gorge at low flow after tower has been extended by one 1.5 m section and walkwire installed.

a

b

c

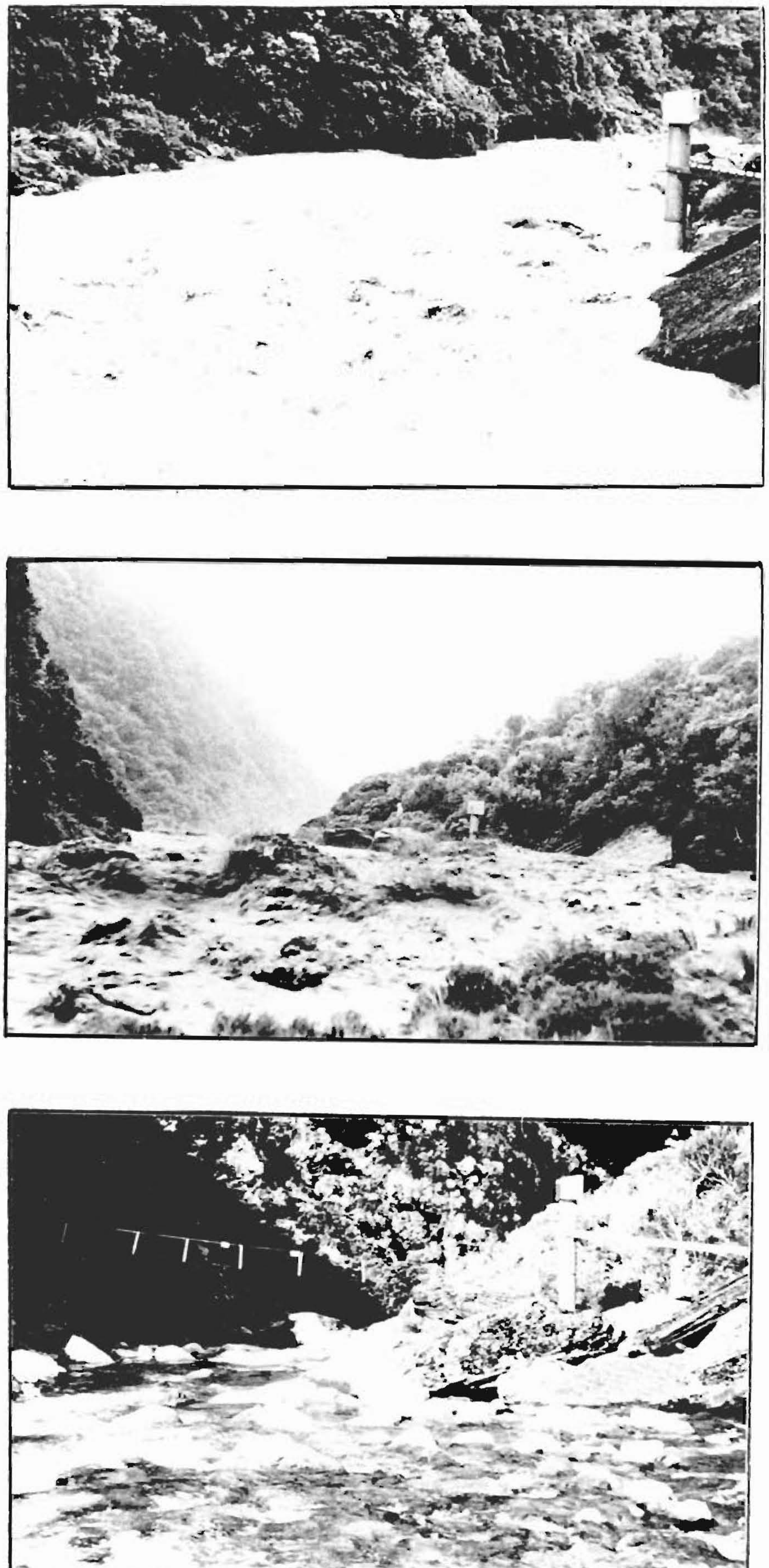


Figure 45: Flow gauging at the Top Cropp site using Gurley current meter on wading rod. The water is colder than it looks.

Seven current meter gaugings on the Cropp at Gorge site and 23 surface float estimates were used to establish the stage height/discharge curve (Fig. 46). The sediment concentration curve was derived using samples taken by wading into the stream to the maximum practicable depth and taking a depth integrated sample using a 500 ml flask. Sediment particles up to 20 mm were sampled from in suspension which would have precluded use of conventional wading rod samplers even if river conditions had permitted it. Suspended sediment concentrations were determined for up to 50 times mean flow. There was no difference observed between rising and falling stage sediment concentrations although some difference may exist (Thompson and Adams 1979). The sediment rating curve (Fig. 47) does not give the maximum sediment concentration for any one flow as the most turbulent part of the flow was not sampled and some loss of sediment occurred at all stages in the treatment of samples (Chapter 10.2). Duplicate samples were taken at each instance, all pairs showed agreement within 15%.

Daily mean flows and sediment discharges were calculated by comparing the rating curves (Figs 46 and 47) with the MWD stage height/flow duration curves in the MWD TIDEDA data analysis system (Thompson and Wrigley 1974). Flood hydrograph data was also retrived from this system.

Results from this study for 1980 are presented in Table 6. The mean annual discharge for 1980 is estimated as 4.63 ± 0.8 m^3 per second. (The estimate takes into account lack of precision in gauging, and uncertainty about the surface velocity/mean velocity ratio.) Conversion of flow to rainfall equivalent, assuming no long-term storage gives 12.0 ± 2.1 metres.

This runoff estimate is in excess of the rainfall value accepted in Chapter 8 of 11 ± 1 metres. This difference is not unexpected as both precipitation from fog and evapotranspiration losses have been omitted from the calculations and more importantly the stage height/discharge relation is very sensitive to minor changes in bed forms in the gauged section. Imprecision in gauging and lack of control on the gauged section make slight overestimation of low flows from the rating curve a distinct possibility.

The maximum discharge observed was 480+ cumecs. This lasted for a very short period and similar stage heights were reached on 24 December 1979 and 24 January 1980. Such short-term discharges may occur with a frequency in the order of one year. This represents an instantaneous water yield of 39.44 cumecs/km^2 which is possibly a New Zealand record.

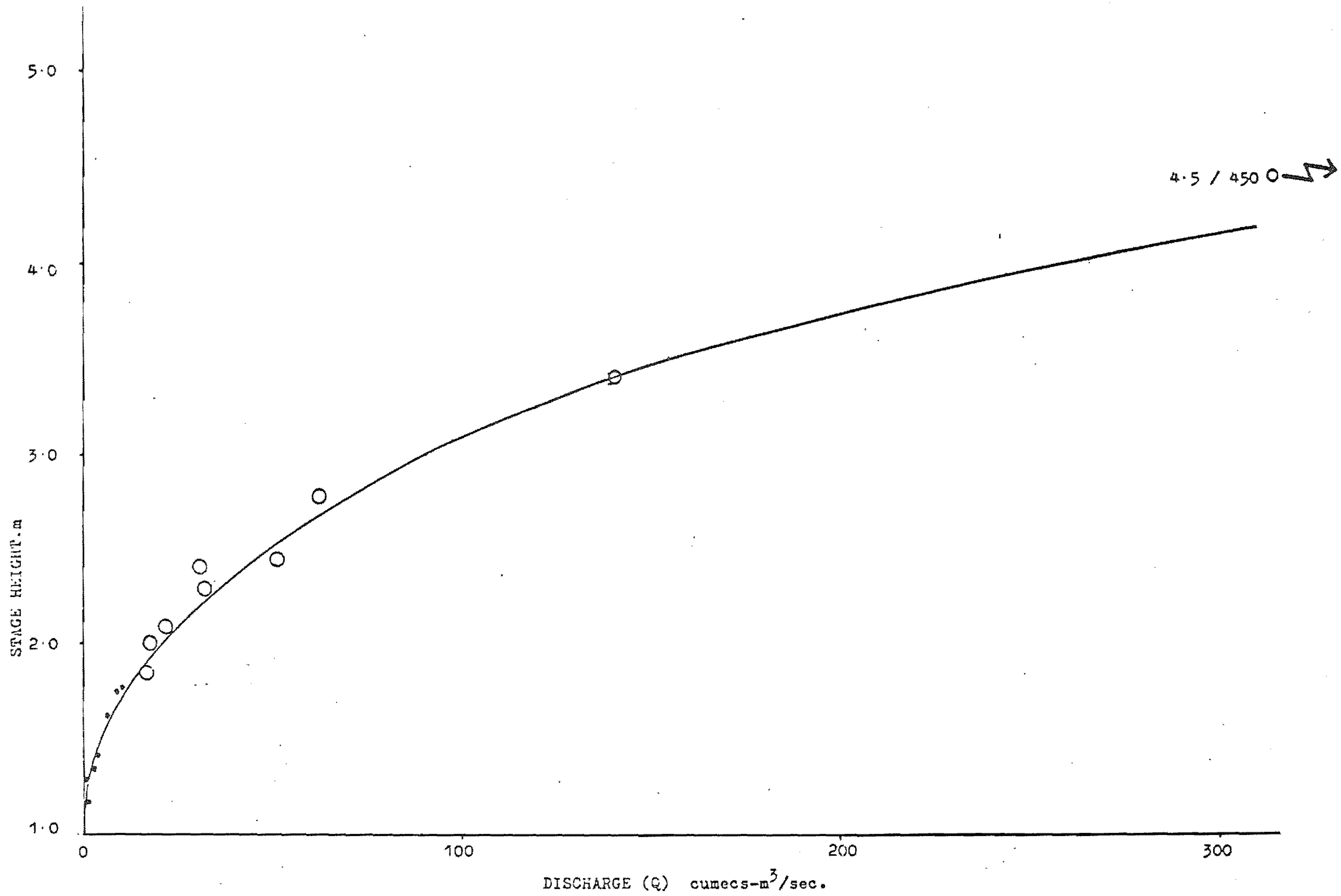


Fig.46:Stage height / discharge rating curve for Cropp at gorge section.Points are current meter guagings,open circles are estimates from maximum surface velocity measurements.

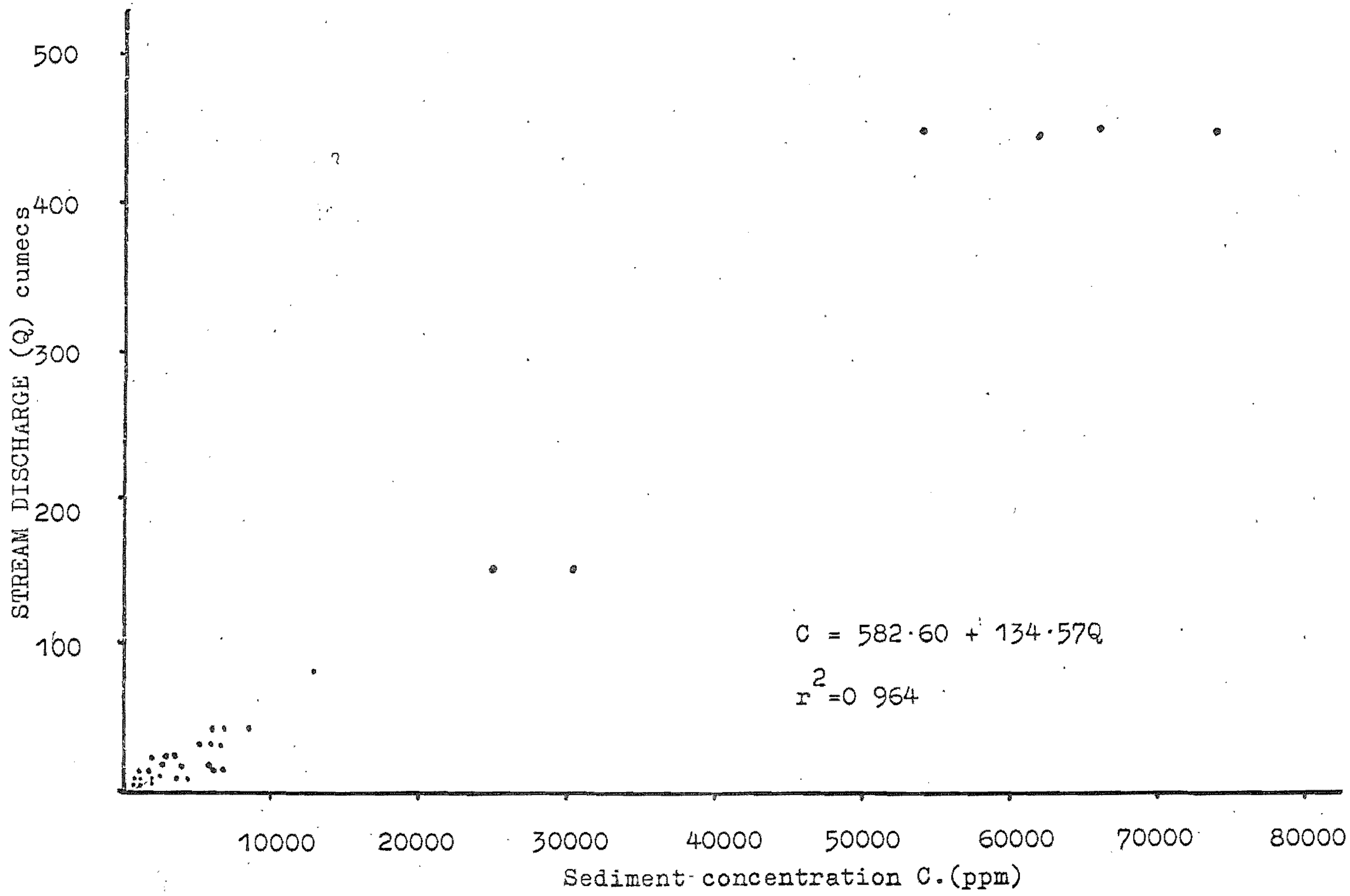


Fig.47: Sediment concentration rating curve.

Low flow gaugings on the tributary streams showed discharges in proportion to catchment area except for Whinging Creek. The low Whinging Creek discharge represents either the minimal storage capacity in the short, steep catchment, or groundwater loss at the gauging station (more likely as the gauged section is on a steep bouldery fan). No significant complications were detected in the discharges of the various sub-catchments under summer low flow conditions. All catchments produced similar sediment concentration values for samples taken on the falling stages of three floods. From the limited evidence collected it seems reasonable to assume that sediment yield is proportional to water yield for the sub-catchments and water yield is proportional to catchment area (Table 6).

9.3 FLOOD HYDROGRAPHS

During 1980 nine flood events of varying magnitudes with discharges up to 50 times the mean flow were observed at the Cropp at gorge site. Of the events which were observed two were considered in detail.

The flood event of the 24 January 1980 produced the highest stage height, discharge and sediment concentrations observed but was short-lived. The flood hydrograph (Fig. 48) shows progressively increasing discharge in response to steady rainfall until 0550 when a 12 minute high intensity rainfall (20.5 mm) was followed within 15 minutes by an extraordinarily rapid rise in river level and, in turn, an equally rapid fall. From this flood peak a steady decrease in discharge accompanied rapidly decreasing rainfall intensity. Considering the fact that there are several peaks in the hydrograph in Fig. 48 each relating to a discrete short rainfall event, application of a standard separation line (Hewlett and Hibbert 1967) shows that 92% of the rainfall was returned as quickflow.

The area under the high peak shown in Fig. 48 represents a flood yield of 288,000 m^3 and an average discharge for 20 minutes of 240 cumecs. The increase in discharge expected from the 12 minute rainfall of 20.5 mm over the 12.17 km^2 catchment is 246,000 m^3 if all is returned as quickflow.

As the rainfall at Cropp Hut is regularly less than rainfall in the upper catchment it is assumed that this rainfall peak represents an immediate quickflow runoff of 20+ mm of rain passing the gorge site within 20 minutes of the rainfall. Close agreement between predicted and measured discharge confirms that the rating curve used to produce the hydrograph is reliable at high flows. If anything, the curve underestimates high discharges. Total

Table 6.

CATCHMENT CHARACTERISTICS

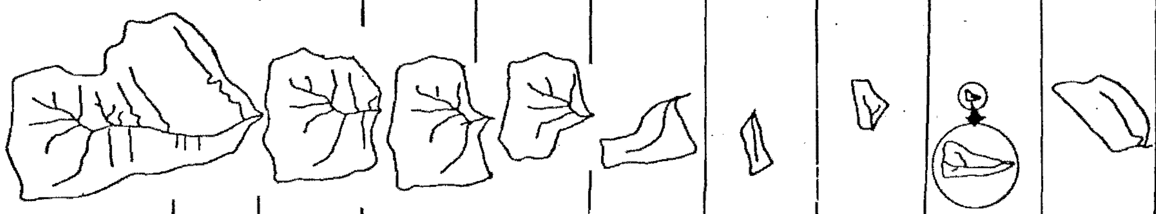
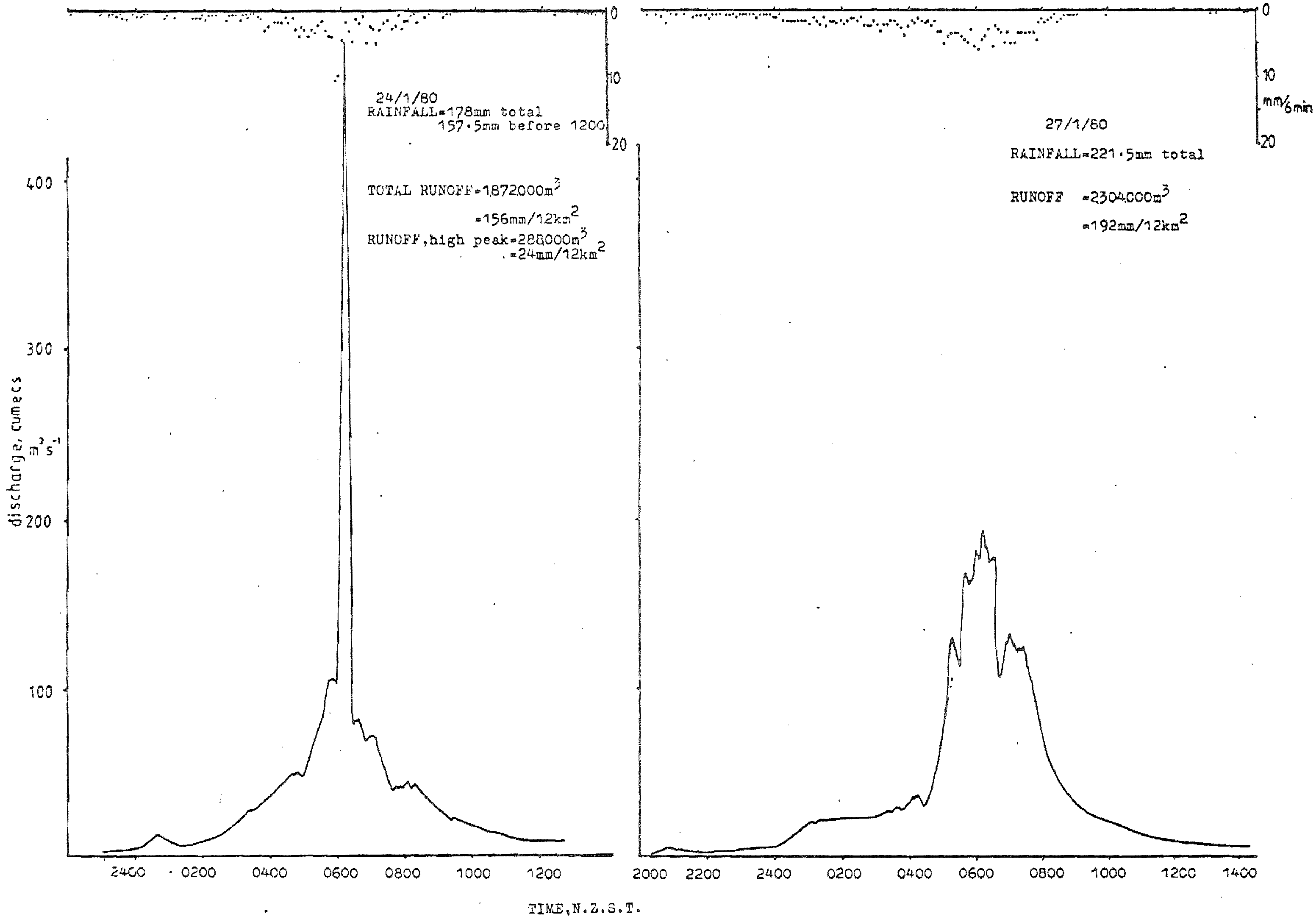
	CROPP RIVER	UPPER CROPP RIVER	MID CROPP RIVER	TOP CROPP RIVER	ROTUNDA CREEK	WHINGING CREEK	DANGER GULLY	LIMPOPO RIVER	NOISEY CREEK
CATCHMENT AREA Km^2	27·0	12·17	10·2	6·13	3·76	6·77	0·92	0·045	3·6
ELEVATION RANGE m	2160 −277	2160 −876	2160 −890	2000 −980	2160 −1240	2000 −885	1500 −880	1320 −1130	2050 −1550
CHANNEL GRADIENT $\frac{\text{fall}}{\text{horizontal distance}}$	21:1	33:1	34:1	36:1	38:1	1;1	34:1	1:1	1:1
SHAPE 									
LOW FLOW DISCHARGE (Q_{min}) cumecs		3·67	2.93	1.70	1,04	0,12	0,27		
DISCHARGE(Q) AS % OF CROPP FLOW		100%	79%	46%	28%	3·3%	7·4%		
$\frac{Q_{min}}{Q_{Cropp}} \times \frac{\text{Cropp Area}}{\text{Area}}$		1	·95	·92	·97	·59	·92		

Fig 48:Hydrographs and hyetographs of floods on 24/1/80 and 27/1/80 as measured at the Cropp gorge recording tower and at Cropp hut.



24/1/80
RAINFALL=178mm total
157·5mm before 1200
TOTAL RUNOFF=1,872,000m³
=156mm/12km²
RUNOFF, high peak=288000m³
=24mm/12km²
27/1/80
RAINFALL=221·5mm total
RUNOFF =2304000m³
=192mm/12km²
mm/6min
discharge, cumecs m³s⁻¹
TIME, N.Z.S.T.

discharge for the storm was equivalent to 154 mm of rain falling on 12.17 km^2 compared with a rainfall measured at Cropp Hut of 157.5 millimetres.

The storm of 27 January 1980 was of a longer duration but lower maximum intensity. The hydrograph for this storm shows three broad peaks following which the stream took six hours to return to mean flow conditions. Again discharge peaks trailed rainfall peaks by less than 15 minutes at the height of the flood. Total runoff predicted by the discharge hydrograph was 2,304,000 m^3 slightly less than that predicted by the rainfall of 2,658,000 m^3 (218 mm). Once again there was general agreement between rainfall and storm runoff estimates. The difference between the two values may represent:

a) Water in storage in the catchment; or
b) Inaccuracy in the river discharge estimate (possibly due to bedform changes during flood flows).

Comparison of the two hydrographs shows the very rapid runoff characteristics of the upper Cropp and the responsiveness of the system to short-term high intensity rainfall events. Much of this quickness of response can be attributed to the large areas of bare rock on the open tops in the catchment. On slopes where soils are established there is other evidence for rapid runoff processes. On many slopes including forest covered areas, runoff during high intensity rainfall events is by surface overland flow which becomes channelised almost immediately in a network of shallow runnels and also by interflow through shallow soil pipes (small subsurface drainage features). In the pelitic schists of the northwest corner of the catchment relatively rapid return flow may occur through the intensely jointed and shattered bedrock.

In recent years interpretation of storm hydrographs on the basis of a classical Hortonian overland flow model has been largely superceded following development of various variable source area models. The Hortonian model (Horton 1940) considers quickflow generation to be a function of surface runoff (overland flow) produced over the entire catchment when rainfall intensity exceeds the infiltration capacity of the soil. The variable source area concept is based on observations of overland flow being restricted to areas adjacent to channels and the idea that the quickflow component of the storm hydrograph is produced from the restricted areas of a catchment where such overland flow occurs. These areas expand with increasing storm intensity and are continually recharged by slower subsurface flow from further upslope (Hayward 1980).

Observed responses of the Cropp River suggest that either widespread Hortonian overland flow occurs producing quickflow runoff of almost all the storm rainfall over the entire catchment (as in the classic model) or that source areas expand to cover most of the catchment very rapidly. Some combination of the two concepts is probably closer to the truth with rapid subsurface "pipe" flow (Ward 1975) combined with overland flow and a greatly expanded drainage net involving small channels contributing to the very rapid discharge of high intensity rainfall from the upper Cropp.

It was intended to compare the monthly and daily mean flows of the Hokitika River at the point where it emerges from its upper valley, with Cropp flows in order to determine relationships between flood intensity and frequency in the Cropp with the Hokitika. This was made difficult due to the Hokitika gauging site having a major siltation problem for three months following the storm of 24 December 1979. The effect of this was that low flow values became vastly exaggerated when using the normal stage discharge relation (Fig. 49). However, the peak flows are probably not too much in error.

Analysis of return frequency of the flood of 24 December 1979 shows it to be a 17 to 20 year event for the Hokitika River. The Cropp record shows that instantaneous discharge never exceeded 140 cumecs (although a very short high flood peak of 400± cumecs occurred between 15 minute punchouts on the Leopold Stephens recorder). However, this flood was of long duration representing 40 hours of continuous high discharge. It was a widespread storm event and involved synchronous high flows in all the tributary streams producing a high peak at the Hokitika gauging site. Although the flood on 27 January 1980 was not so long-lived it again involved similar intensity in the Cropp and also produced a high discharge in the Hokitika. The flood on the 24 January 1980 however, gave a record discharge in the Cropp (Fig. 50) as a result of a short-term high intensity rainfall event but was of little significance at Hokitika, presumably because of the short duration. Thus it would seem that high instantaneous discharges in the Cropp are dependent on high intensity rain whereas in the Hokitika they depend on high total rainfall and long enough duration for all sub-catchments to be contributing high discharges at one time.

Fig 49:Uncorrected discharge estimates for the Hokitika River at Colliers gorge for early 1980 showing the effects of siltation of the rated section.

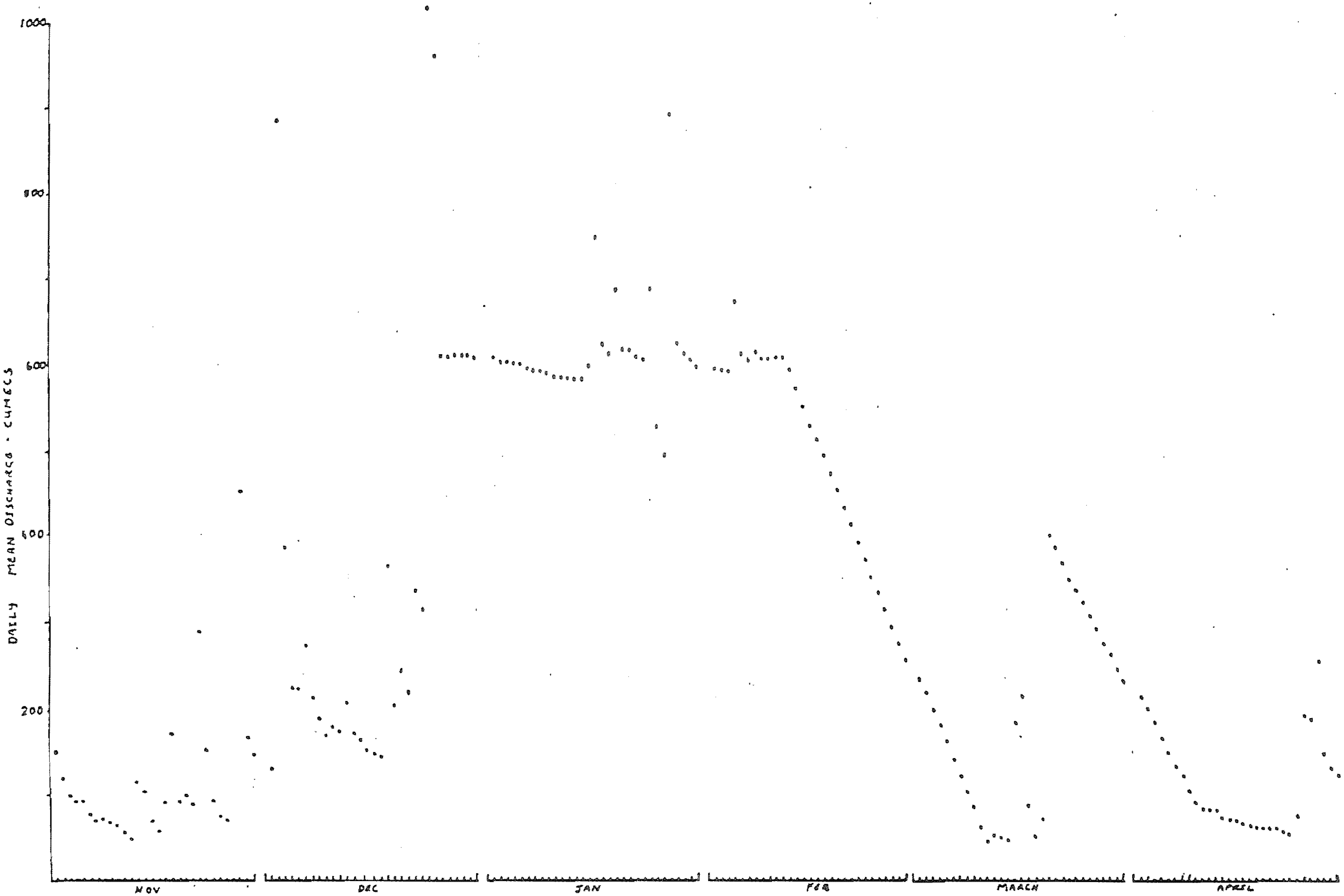
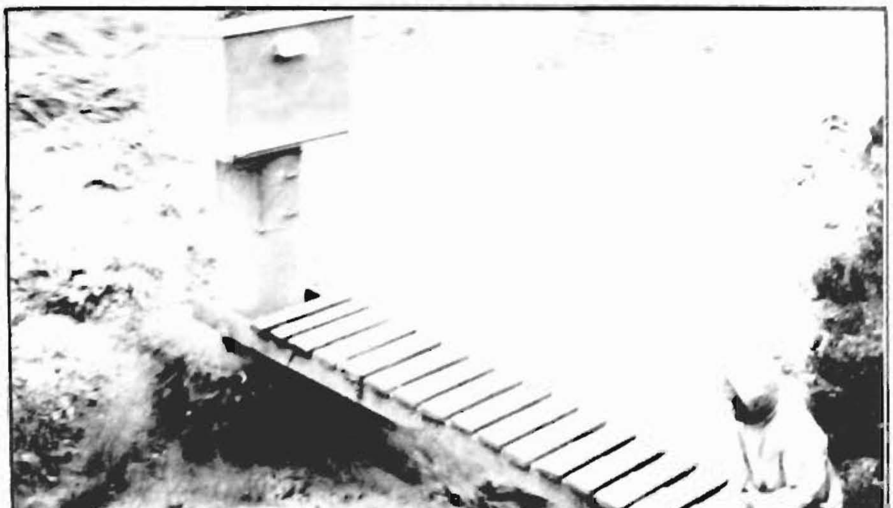

1000
800
600
400
200
DAILY MEAN DISCHARGE - CUMECS
NOV
DEC
JAN
FEB
MARCH
APRIL

Figure 50: Flood of 24 January 1980. At maximum stage water was thigh deep on river flats and just covered the bridge to the gauging tower.

a

b

c



CHAPTER TEN: SEDIMENT YIELD STUDIES

10.1 INTRODUCTION

The main aims of the study of sediment yield were: to estimate the amount and type of sediment transported out of the upper Cropp area; to identify the various sources of sediment within the catchment; and the areas of high sediment yield. Ultimately it was hoped to observe patterns relating sediment yield to climatic or geologic controls.

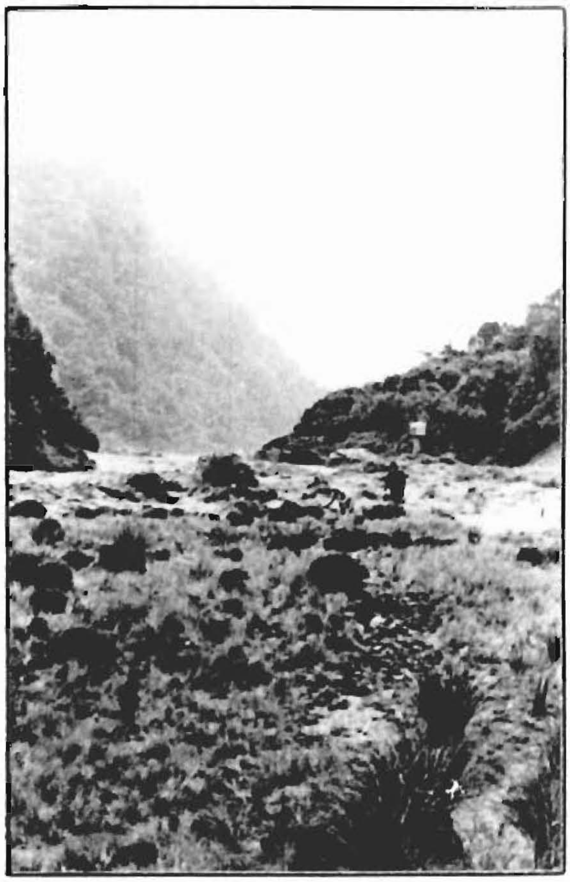
To do this, records of suspended sediment discharge defining a sediment rating curve were obtained for the Cropp gorge site. Combining stage height records from the MWD water level recorder at the Cropp Gorge with the discharge and sediment rating curves allows estimation of the total suspended sediment discharge for 1980. Bedload yield estimates were made for individual storm events affecting a small (4.5 ha) catchment using a small debris dam. Relative contribution of bedload by tributary streams was investigated using a natural trace dilution method (Martin 1972). The different lithologies present in the gravels were used as indicators of gravel source areas. The simulated effect of abrasion on Cropp River bed material was investigated in a mechanical tumbler.

10.2 SUSPENDED SEDIMENT

Because of the lack of a suitable bridge from which to sample suspended sediments in the Cropp at gorge site until October 1980, after which no significant floods were measured, suspended sediment samples were taken using non-standard techniques and without a standard sampler. Samples were taken using 500 ml flasks, by wading into the flooded river as far as was practicable and taking a depth integrated sample (Fig. 51). The sample was rejected if the flask was overfilled. The grain size of suspended material was too coarse (up to 20 mm+ blade of schist) to allow standard samplers to be used.

Samples were vacuum filtered at the Cropp Hut, oven dried and sealed in plastic bags for weighing on return to Christchurch. Because there were no laboratory facilities at the Cropp Hut the precision of sample treatment was affected. However, all inaccuracies involved losses of material (of fines during filtering and spillages) so concentrations obtained are minimum values.

Figure 51: Suspended sediment sampling.



By plotting the sediment concentration against the discharge at the time the sample was taken (from the stage height/discharge curve (Fig. 46)) a sediment concentration rating (Fig. 47) was derived.

$$C = 582.60 + 734.57\ Q$$

with a correlation coefficient $r^2 = 0.964$

where C = sediment concentration ppm

Q = flow in m^3/sec.

Using this relation sediment yield estimates for 1980 were made. This gave a mean annual sediment yield of 1080 tonnes per day with a maximum daily yield of 22,943 tonnes on 25 August and a maximum instantaneous discharge of 33.6 tonnes per second on 24 January.

The mean value is equivalent to a suspended sediment yield of $32{,}410 \pm 11{,}290$ tonnes/km^2 per year (at the 95% confidence level). This figure is broadly compatible with values for specific sediment yield of the Hokitika, 17,070 tonnes/km^2 per year (Griffiths 1979) and 21,000 (J Adams 1979); for the Ivory Glacier at 12,000 to 25,000 tonnes/km^2 per year (McSaveney *et al.*, in press) and with the predictive equation of McSaveney *et al.* in Fig. 52.

From presently available data it cannot be determined just how representative this 1980 yield figure is. However, the climatic and stream flow studies did not suggest that 1980 was particularly unusual in the Cropp.

There was no evidence of any clearly defined loop rating (a loop rating is produced where, during a flood "the available load is exhausted early in the flood and exhaustion causes . . . more load to be carried during rising then falling river flow."; Thompson and Adams 1979) in the sediment discharge data although such a relationship may not have been detected given the limited time series data.

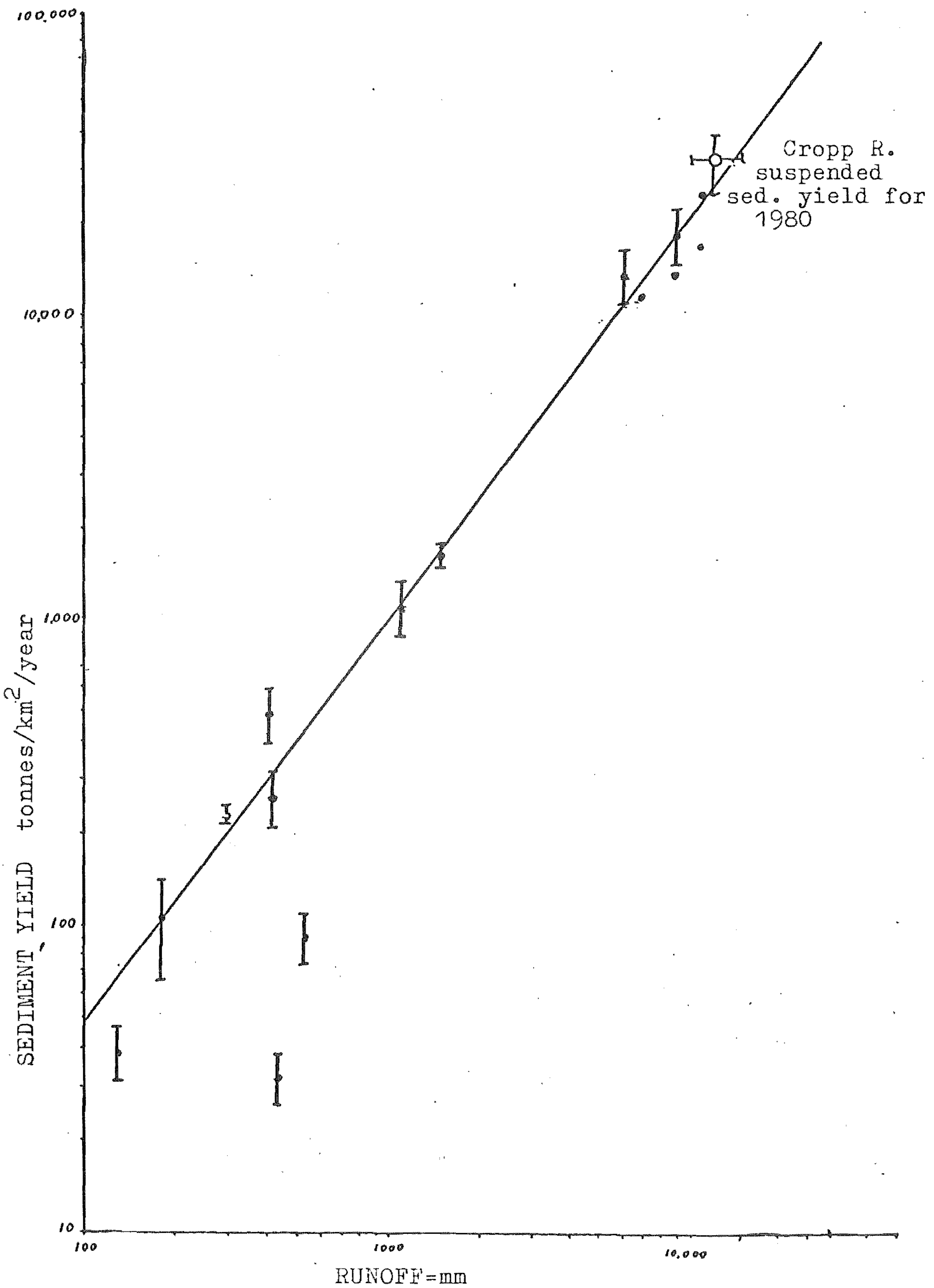
Suspended sediment data collected for tributary streams of the upper Cropp were all at moderate flows on falling stages and gave similar concentration values. From the data available there were no indications of locally active sources of suspended sediment dominating sediment supply.

10.3 BEDLOAD

Estimation of the contribution of bedload material to total sediment yield is difficult especially in a steep mountain stream like the Cropp. In the Cropp there are no unfilled reservoirs which could act as sediment traps.

Fig 52: Relation between sediment yield and annual runoff from McSaveney *et al* (in press) with the upper Cropp suspended sediment yield for 1980 indicated.

FIG 52.


100,000
10,000
1,000
100
10
SEDIMENT YIELD tonnes/km2/year
100
1000
10,000
RUNOFF=mm
Cropp R.
suspended
sed. yield for
1980

Construction of large structures for sampling bedload would be impractical and the requirements for long-term sampling to derive a meaningful rating curve if bedload could be sampled could not be met within the scope of this project.

McSaveney (1978) suggests that the magnitude of suspended sediment yield from the high rainfall area at the Southern Alps, especially in schist terrain makes high bedload yields unlikely. As a result no effort was made to measure bedload yields from the Cropp. Instead the bedload sediment train was followed by detailed observation of sediment type and size within stream channels. The type and amount of sediment moving in a small high altitude catchment was observed using a small debris dam.

Data from these studies combined with observations on changes in bed morphology during floods, observations of catchment conditions and experimental investigation of the effectiveness of abrasion in changing bed material characteristics allows a qualitative assessment of the magnitude and type of bed material movement that occurs.

Debris Dam

In order to examine slope and stream processes in a small, steep catchment more closely, a debris dam was built in the Limpopo River downstream of seven sites monitored for mass movement (see Chapter 10.4). The catchment area above the dam was 4.5 ha and had an elevation range of 200 metres. The stream channel was deeply incised into bedrock for half of its length and was otherwise a boulder bed torrent.

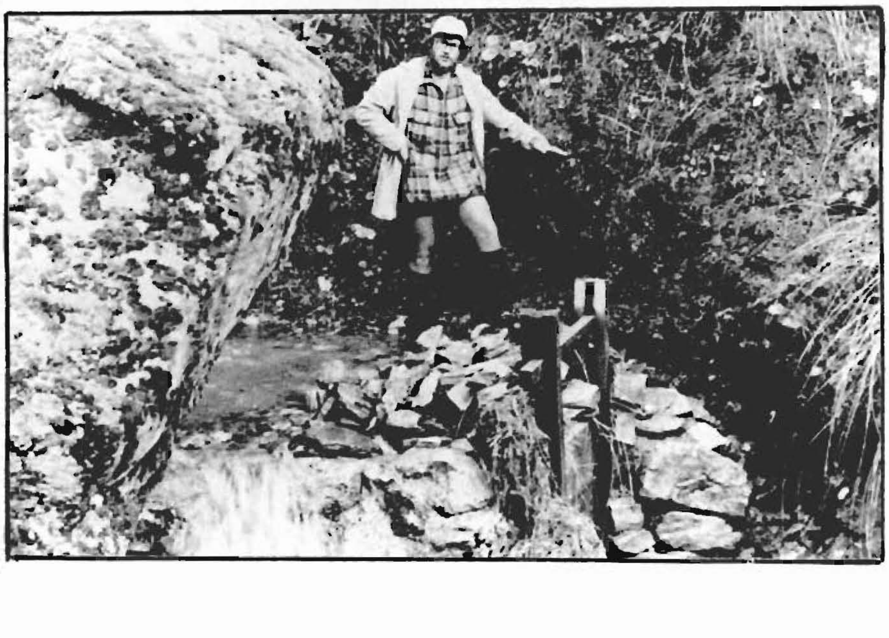
The dam comprised a one metre concrete and boulder structure with a removable sluice gate which allowed the reservoir to be emptied (Fig. 53). It was constructed on a bedrock base at the top of a steep cascade. At the upstream end of the reservoir a two metre drop over bedrock formed a plunge pool up to 50 cm deep when the dam was full. A 5 cm gap was left between the bottom two boards to allow free drainage at low flows. This gap may have allowed some sediment to pass through the dam although this was not significant at high sediment discharges.

The dam was inspected and emptied after each flood event. The surface profile of the sediments contained by the dam was measured after each flood enabling the volume and weight of sediment to be computed (Fig. 54), using an assumed density of 2.2 tonnes/m^3.

Figure 53: Debris dam in Limpopo River (a) showing removable sluice gate; (b) Dam overfilled with armoured bed of coarse chert boulders, maximum water level overflowing dam is indicated.

a

b



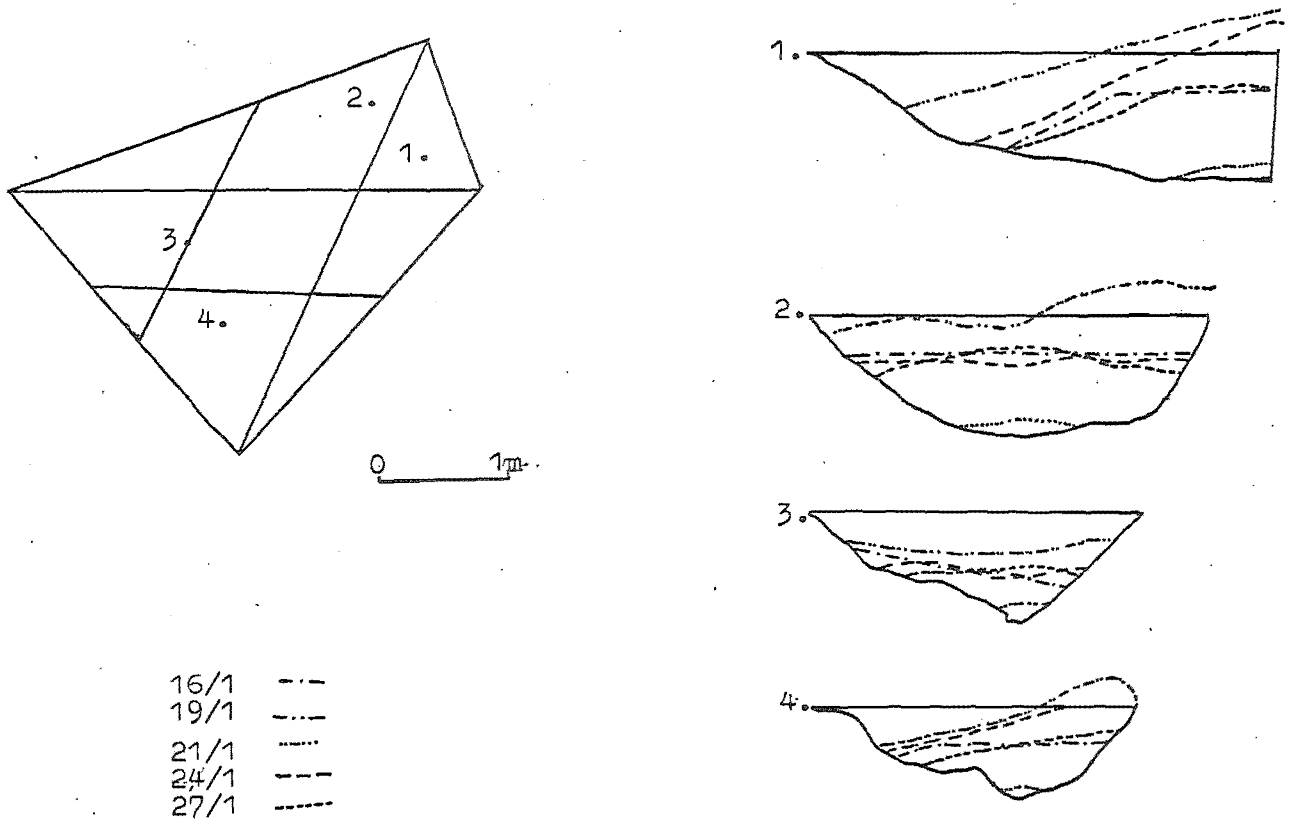


Fig.54:Profiles of sediment trapped in debris dam during storm events of January 1980.

In all, debris from five events was trapped in the reservoir while it was being monitored (during January 1980). The reservoir was found to have overflowed and was emptied on two further occasions later in the year, and found not full once. The total volume removed from the dam during 1980 was 15+m^3. (33+ tonnes at 2.2 tonnes/m^3). From a catchment area of 4.5 ha this represents a minimum sediment yield of 733 tonnes/km^2 per year. This value represents bedload only (no sediment finer than granule size was trapped in the dam) and is an absolute minimum as the dam overflowed and passed sediment through on four occasions. This included six months over winter when the dam was not emptied. For part of this time the dam site and upper catchment were completely snow-covered.

As shown in Fig. 55 sediment trapping is more affected by short-term rainfall intensity than by total rain during a storm event. The Limpopo probably has runoff characteristics with response times in terms of minutes so a 12 minute rainfall of 17.5 mm on 19 January 1980 would have produced a greater instantaneous discharge than was produced in the less intense storms of 15 and 16 January 1980 and so produced a higher sediment yield (Fig. 55).

The more intense and larger storms of 24 and 27 January 1980 involved deposition of less material than the 19 January 1980. This reflected a combination of scouring of material out of the dam during prolonged high flow and possibly a lack of sediment availability.

Lack of sediment availability relates to in-channel sediments only. During the flood of 19 January 1980 material trapped in the debris dam included shards of painted schist which had moved from a bedrock face 200 m away during that event. Indeed, most of the sediment trapped by the dam in any of these events (50%+) consisted of small shards of schist which appeared to have moved from slopes into channels in the one event. It is unlikely that supply of sediment from this source would be depleted as the bedrock in this area is highly fissile and tectonically shattered to considerable depths.

Trapping efficiency of the debris dam was low either at low flows or where sediment volumes in the dam exceeded about 2.25 m^3. At low flows material up to 5 cm was able to pass through the base of the sluice gate so that very small events were not monitored. At high discharges the reservoir became brim-full.

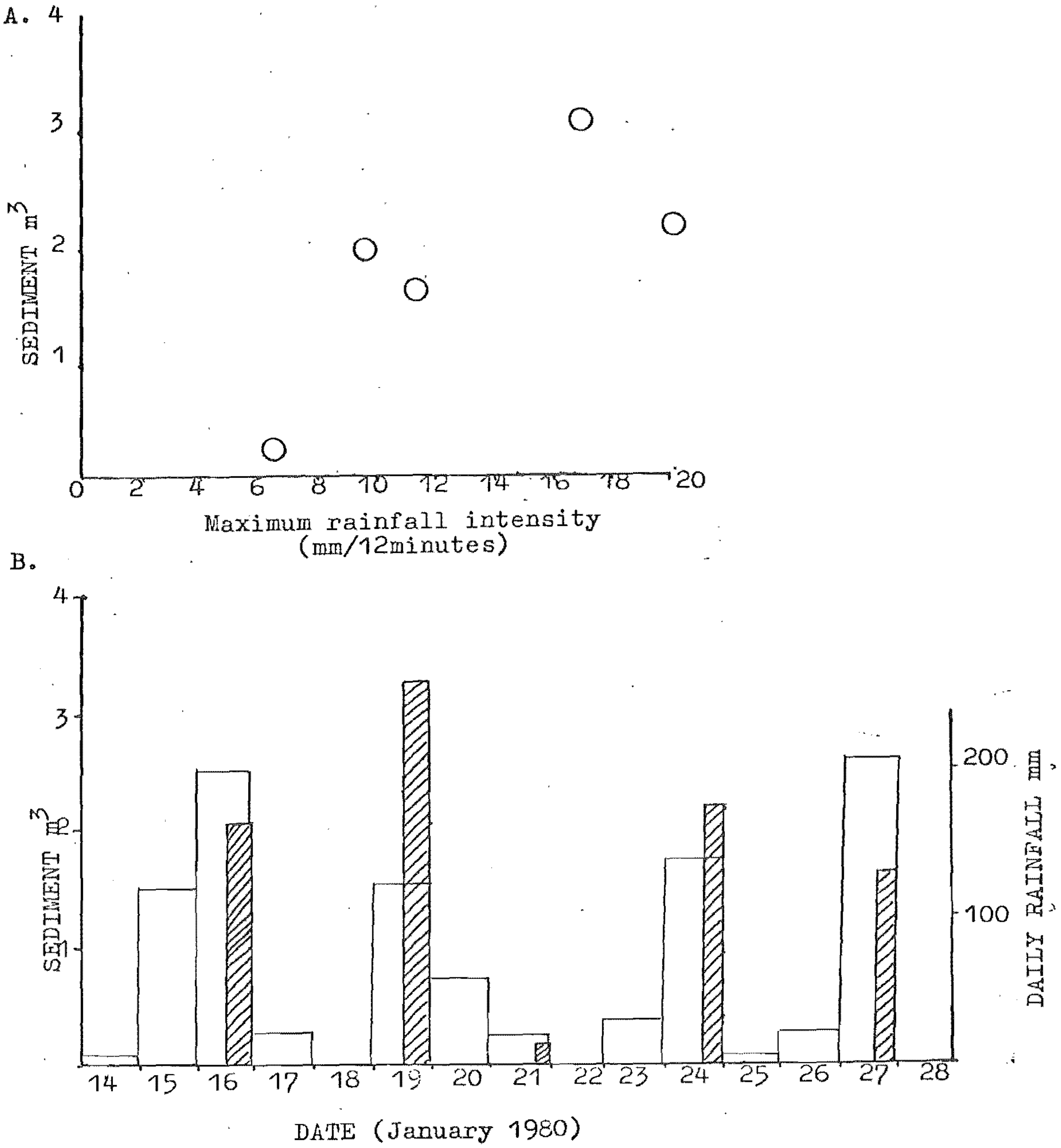


Fig.55.Volume of material trapped compared with the maximum rainfall intensity for each storm event(A.) and with the total daily rainfall(B.).

Eventually armouring developed on the surface of the debris and all sediment was passed over the dam.

Determination of the size and siting of the dam was largely arbitrary, relating most to availability of suitable construction materials and sites. The dam was considered appropriate in the light of studies such as those of O'Louglin et al. (1978) on larger catchments using similar-sized traps, and because of the prediction by McSaveney that bedload was likely to be a minor contribution to sediment yield in tyis type of mountain catchment situation.

Observations of sediment movement in the Limpopo suggests that large-scale sediment movement occurs once stream flow exceeds a critical value. Most sediment which moved into the stream channel was in the fine gravel-size range and this passed from the rock face out of the catchment within single events (as discussed previously). The bed of the stream was composed almost entirely of coarse chert boulders, although chert comprised only a minor proportion of the bedrock in the catchment.

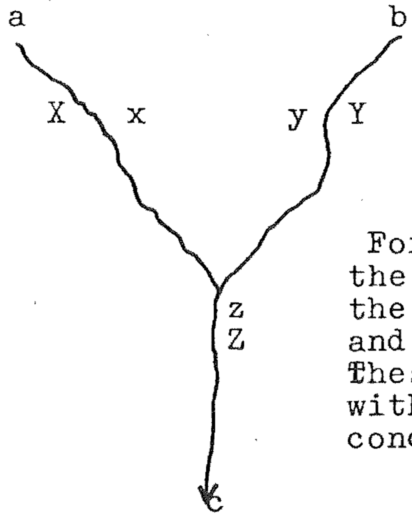
Coarse detritus within the stream channel was limited in supply. Most of the bedload sediment is of fine gravel-size and is derived from exposed bedrock and talus which is present in almost inexhaustible supply in this very steep, small catchment.

Tracer Dilution, Bed Material Study

In order to determine the relative bed material discharges from the four main sub-catchments of the upper Cropp River into the main stream, a technique known as natural tracer dilution (Martin 1972) was applied. The technique is based on the concept that below the confluence of two streams the composition of bed material will be related to the composition of the bed material in each contributing stream multiplied by the streams sediment discharge. Thus if the composition of the bed material in the two contributing streams and the main stream are known (and are significantly different) the relative bedload discharges can be calculated (see Fig. 56).

This technique relies on there being a distinctive component in the bed material which acts as a tracer, that it can be readily identified, and which is hydraulically indistinguishable from the rest of the bedload. Martin (1972) used the chert component of a greywacke-dominated gravel to study the contributing catchments in Kowhai Stream.

Fig 56: Basis for estimates of relative bedload yield of contributing streams.



For streams above the confluence the concentration of a component of the bed material is x and y respectivel; and sediment discharges are X and Y. These combine to form a stream with sediment discharge Zand sediment concentration z.

The constant relations assumed for this system are:

$$Zz = Xx + Yy$$

and $Z = X + Y$

Therefore: $(X+Y)z = Xx + Yy$

$$Xz + Yz = Xx + Yy$$

$$Xz + Xx = Yy - Yz$$

$$X(z-x) = Y(y-z)$$

$$X = Y\left(\frac{y-z}{z-x}\right)$$

$$\frac{X}{Y} = \frac{y-z}{z-x}$$

Thus if the concentration of a component of the normal sediment load is known for streams above and below a stream confluence the relative contribution of each tributary stream can be calculated.

In the Cropp the sediments are of a more varied composition with six readily identifiable lithologic types present. Thus mixing proportions of different lithologic tracers provided an internal check on the values determined. Three of the components were present in quantities too small to give valid results.

Assumptions made were: that all sediments below the confluence were from the contributing streams; that complete mixing had taken place; that tracer input, transport and abrasion rates were all identical to rates for the bulk of the sediment. (As the results show, these assumptions do not always hold.)

Between 100 and 150 rocks were counted for tributary and main streams both above and below each of three stream confluences. The sediment of particular interest was the material which commonly moved as bedload in flood conditions in the Cropp. Hence, sampling was restricted to within 3 m of active or flood channels and to particles of 1 to 40 cm maximum diameter.

Because collection of bulk samples large enough to be statistically representative of coarse river sediment was deemed impractical in the Cropp a method of randomising the counting and measuring of sediment was employed. This is a variation of the modified Wolman method for sampling alluvial gravels described by O'Loughlin (1968). A 1.5 m long stick marked in 10 cm sections was thrown onto the area of gravel to be sampled and the clast adjacent to each 10 cm mark was measured and classified.

Results of sampling at three stream junctions are presented in Table 7 and the results of comparison of relative contribution of traces is presented in Table 7A. The minor components in each case were present in quantities too small to give statistically significant results.

Of the three stream junction systems studied, only the Rotunda Creek-upper Cropp confluence produced consistent results. The others showed effects of comminution (removing the more friable pelitic schists) of differences in hydraulic behaviour between chert and other bed components or of sampling error where the rocks sampled were not representative of the bedload either because stable lag deposits were sampled or samples were not from the active channel.

Negative values of relative sediment discharge on three of the component tests, resulted from their being more or less of the component present below the confluence than in either of the contributing channels. This situation

TABLE 7

SITE	CHERT	PELITIC SCHIST	PSAMMITIC SCHIST	GREENSCHIST	MARBLE	QUARTZ
DANGER GULLY	58·7	7·5	17·5	15	–	1·25
CROPP ABOVE DANGER GULLY	29	49	5	15	2	–
CROPP BELOW DANGER GULLY	38	30·4	19·5	8·7	2·17	1·1
WHINGING CK	20	72·5	2·5	–	–	3·75
CROPP ABOVE WHINGING CK	22·5	55	8·75	11·25	–	2·5
CROPP BELOW WHINGING CK	29	49	5	15	2	–
TOP CROPP	28·1	43·75	13·5	11·5	2·1	1·04
ROTUNDA CK	3·75	33·75	43·5	13·75	–	1·25
MID CROPP	14·3	38·7	30·9	8·7	1·6	0·8

TABLE 7b

ROTUNDA CREEK

ROTUNDA CREEK	q_1/q_2	% contribution to Cropp River sediments
Chert	·765	43
Pelitic sch.	·98	49·5
Psammitic sch	·724	48
WHINGING CK		
Chert	-0·72	–
Pelitic sch	-0·255	–
Psammitic sch	0·66	39
DANGER GULLY		
Chert	0·43	30·2
Pelitic sch	0·81	45
Psammitic sch	-6·25	–

%contribution is the sediment discharge of the contributing stream implied by the sediment composition as a percentage of the sediment discharge of the Cropp R.

could not have existed if the initial conditions were all fulfilled and must have resulted from either enrichment of a particular bed material component by abrasion of other components (ie chert compared with pelitic schist for Whinging Creek) or large-scale abrasion of individual components over short distances.

Results from Danger Gully suggest that the contribution of chert to the main Cropp was significantly less than the contribution of pelitic schist. This is because chert in Danger Gully forms a coarse resistant lag deposit whereas pelitic schist dominates in the finer, more mobile bed material fractions. That pelitic schist dominates the mobile bed material in this area, is shown by the composition of material trapped in the debris dam in the Limpopo River where almost all trapped material was pelitic schist, Whereas bed-forming elements were almost entirely chert. This is, in part, at least, due to the susceptibility of pelitic schist to comminution and transportation as saltation or suspended load.

As they enter the Cropp, Whinging Creek sediments do not significantly change the composition of sediments in the main valley because of the small catchment area from which they originate. Although the psammitic schist values suggest a 39% contribution from Whinging Creek to the Cropp total, this is an unreasonably high value for a small, relatively stable catchment. The values for the Cropp River above and below Whinging Creek show significant differences which cannot be attributed to contributions from Whinging Creek. Chert is enriched by 6.5% of total clast freqneucy and pelite depleted by 6% over the 200 m of channel between the two Cropp samples. This may reflect selective comminution of pelitic schist.

Values of relative contribution for the major sediment components at the Rotunda Creek confluence show a consistent trend with the Top Cropp contributing 50.5 to 58% of bed material clasts. The small value for pelitic schist contribution from the Top Cropp is probably due to the close proximity of pelitic schist sources in the Top Cropp (20 m from the sampling site) and their relative immaturity (angularity and low roundness of sediments closer to the source) resulting in greater abrasion of pelites in the Top Cropp than in Rotunda Creek.

A ratio of 1:1.3 for Rotunda Creek:Top Cropp contribution of sediment is indicated by these results. When compared with the ratio of catchment areas (1:1.63) and low flow discharges (1:1.64) the relative sediment discharge value suggests that Rotunda Creek contributes more bedload sediment per unit area than the Top Cropp. This may be due to a lack of coarse sediment sources in the glacial U-shaped valley of the Top Cropp. The geomorphology of Rotunda Creek is almost entirely fluvial including relict fan surfaces. This contrasts with the upper Cropp which has large areas of glacial surfaces that do not contribute coarse detritus to the stream bed.

This study failed to identify areas of major sediment contribution. However, it does suggest that significant comminution of unstable bed material could occur over very short distances within the Cropp systems. Data from the Top Cropp suggests that relict glacial surfaces are not active in terms of coarse sediment supply to the stream system.

That comminution is effective in converting bedload to fine-grained suspended load is also suggested by observations of sediment compositions in the lower Cropp where a range of ultramafic rocks are present as tracers. At the confluence of the Cropp and Whitcombe Rivers greater than 50% of the large bed material is locally derived ultramafic and related rock. At the head of the lower gorge 1 km away (and 250 m higher) 80+% of bed material is quartzofeldspathic schist. As there is no significant storage of bed material in the gorge it is likely that much of the bed material entering the gorge is converted to suspended sediment and is not deposited at the change in stream gradient at the Cropp-Whitcombe confluence. It is possible that under normal circumstances only negligible amounts of bedload are transported the entire length of the Cropp. Certainly no metapelite is transported for long distances in the Cropp as bedload.

Abrasion Testing

In order to test the hypothesis that abrasion of gravel occurs in sufficiently large amounts to produce changes in stream sediment composition over short channel reaches in the Cropp River and its tributaries, gravel samples were subjected to artificial abrasion, and grain size distribution changes were noted.

Adams (1979b) suggests that schist pebbles decrease in volume at an exponential rate with distance from the river headwaters and that the susceptibility to abrasion of gravel is higher near the river source because of a higher proportion of unsound pebbles than in lower reaches of the river. Such a mechanism is suggested by the decrease in pelitic schist relative to other rock types observed along reaches of the Cropp River. If selective abrasion of unsound pebbles is the mechanism for changing grain size and composition of a gravel population along a stream, it should be possible to show a decreasing ability to abrade gravel further from its source.

To do this, six 2 kg samples of bed material (smaller than -6∅) were taken from the Limpopo River, Danger Gully, and the Cropp River. These were first experimentally abraded for 10 minutes, then for a 2½ hour period in 20 cm metal tumblers (Franklin and Chandra 1972) at 60 rpm. Material less than 2.00 mm was lost from the wire mesh tumbler to the surrounding water tank. This is probably a very poor simultion of transport of gravel in a high country river system. However, it does provide a fair test of the relative abrasion susceptibility of gravel samples.

The results of this study, presented in Fig. 57, showed a consistent grain size reduction in all samples. There was, however, no increase in abrasion for near source gravels or for gravels with a high percentage of pelitic schist (samples 01, 02).

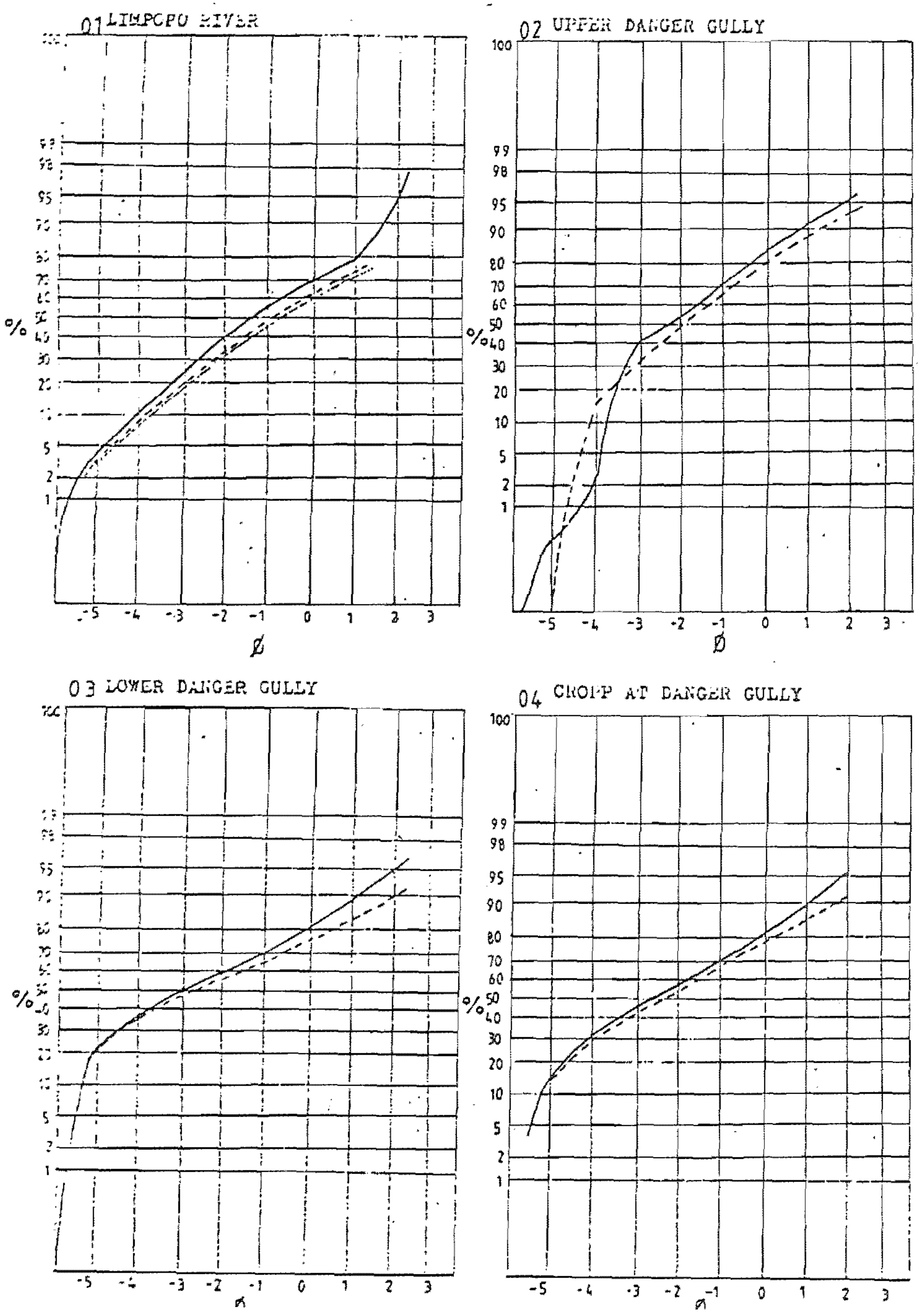
The grain size change was greatest for fine gravel sizes and least for coarse gravels, except for sample 02 which shows the effects of one pebble changing from the -5 ∅ to the -4 ∅ size class.

One sample showed considerably more change than the others; 06 showed a 0.6∅ fining in the fine gravel classes. This large movement may have been due to the large percentage of coarse material (16% greater than -5 ∅) in the sample, as discussed below.

Observations of change within the shingle samples are limited. Pebbles larger than -2 ∅ showed a rounding effect with a loss of jagged edges. There was no selective abrasion of different lithologies noted.

Interpretation of these results is difficult as many parameters are involved in the experimental system which is a very incomplete model of the natural system. Marshall (1927) suggests that grinding is by far the most rapid mechanism of grain size reduction in tumbler experiments. Impact is significant if the grain size range in the sample is large and abrasion is by far the slowest mechanism of grain size reduction. Thus grain size can be expected to change most rapidly in a poorly sorted sample with significant fines.

Adams (1978b) determined abrasion coefficients for Alpine Schist pebbles (and a large range of other lithologies), from experimental studies where individual pebbles were abraded under standard conditions using the Sternberg Law. Adams (1979b) concludes from a simple study of gravels of the rather complex Clutha River system, that the abrasion coefficient of a gravel population increases upstream with an increasing proportion of "unsound" pebbles. Results from this experiment give rates of grain size



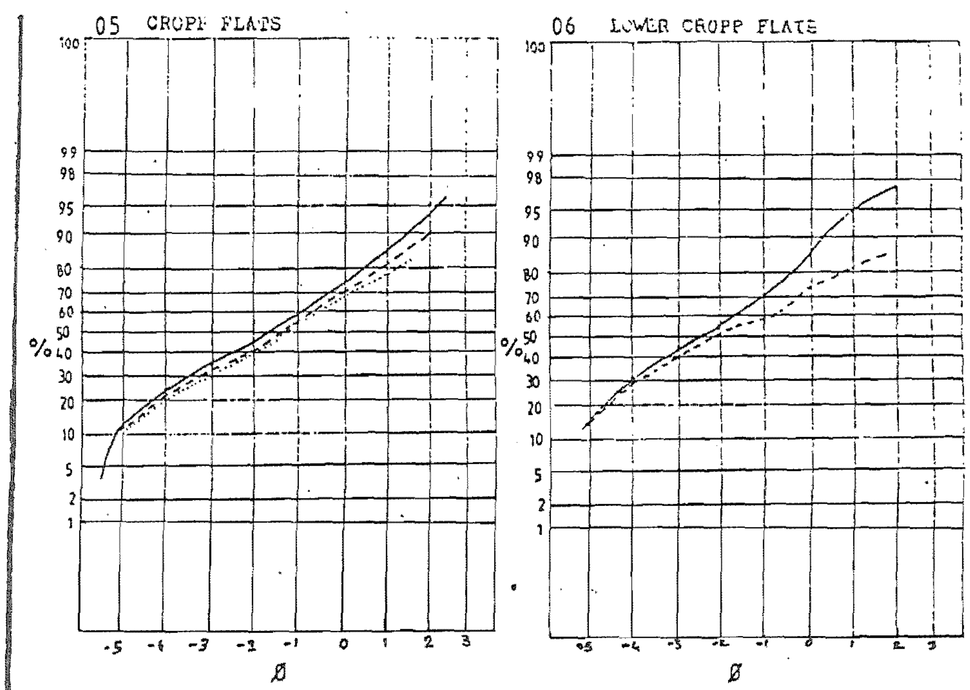


Fig.57:Cumulative grain size distribution curves of gravel samples from the Cropp River showing changes in grain size after experimental abrasion for ten minutes(- - - - -) and two and a half hours(··········).Increasing sample number reflects increasing transport distance(downstream distance of sample locality from sediment scources).

reduction greater than the Adams (1978b) coefficient predicts, yet shows no relation with distance from the river headwater as would be expected from Adams (1979b).

It is possible that the grain size changes observed in the experiment and by Adams (1979b) reflect different grain size distributions in the gravels of a river system further from the headwater and different comminution processes active as a result. If abrasion is less important than grinding and impact and these mechanisms are dependent on grain size distribution it is possible that the changes observed by Adams (1979b) are an effect of increased sorting of gravels downstream and an increasing importance of abrasion relative to impact and grinding as a result.

The experiment simulated travel in the order of 2.5 to 3 km (for 2½ hours) and yet produced no visibly conspicuous change in sorting or composition and only minor grain size changes. This is in contrast with observations in the Cropp catchment where changes in gravel size and composition (especially the amoun t of pelitic schist) are often rapid and systematic over tens of metres of river channel. This suggests that the abrasion simulation model is not a good one and that tumbler experiments are best compared with other tumbler experiments. It should be possible to model grain size reduction as it occurs in rivers like the Cropp by experimentally identifying the processes which cause the reduction and by scaling the experiments to the river system. This would require considerable effort with limited chances of success as the mechanics of sediment movement in rivers like the Cropp are still poorly understood.

Observations on Bedload Movement

Direct observations of flood bed forms in steep mountain streams in New Zealand are restricted to studies on the Torlesse Stream catchment (see Hayward 1980). In this study sediment supply is limited because of the nature of the catchment and was observed moving as slugs or waves probably relating to individual mass movement or sediment supply events. Hayward does not describe events involving supercritical flow or sediment movement episodes which disrupt the main channel structures of the Torlesse Stream.

Observations in the Cropp River relevant to interpretation of the amount and type of bedload movement include:

1 Sediment trapped in the debris dam was largely finer than 4 Ø and was mainly pelitic schist, a lithology under-represented in "armoured" bed sediments;

2 The Limpopo River regularly transports boulders weighing 20 kg or more;

3 Small flood events, with recurrence intervals of weeks or months and discharges up to 10 times mean flow, transported bed material of up to 15 cm (long dimension) in the Cropp and lower tributaries, but only for short distances (stripes painted on river gravels were only partially dissipated during two months with only minor floods).

4 During the flood event of 24 January 1980 aggradation of the Cropp River bed occurred over much of its length (Fig. 58) and flow velocities were very high during the flood maximum (maximum surface velocity of 7.5 ms^{-1}). Flow was supercritical (Froude number of 1.5$\pm$) and bed roughness effects were negligible (manning n approximately 0.015). All major bed roughness elements were buried by aggradation during this event as was the pool and riffle structure of parts of the upper Cropp (Fig. 59).

5 Large boulders (in one instance a boulder 2 m x 1.5 m x 1 m) were seen to have moved tens of metres during single events in the Cropp River (Fig. 60).

6 During the large storm event (24 January 1980) when aggradation occurred in the Cropp River, tributary streams also underwent significant aggradation to depths of greater than 2 metres.

From observations made it is proposed that bedload transport in the Cropp follows a pattern of minor saltation and traction movement of fine-medium sized material during small-moderate flood events with relatively little total bedload movement. In upper catchments considerable movement of fine gravel occurs much of which is reduced to suspended sediment sizes by the time it gets to the main channel by processes of impact, grinding and abrasion.

During infrequent large events an influx of sediment from many tributaries combined with high flow velocities cause bed aggradation which swamps large bed roughness elements allowing further increase in flow velocity (Simons _et al._ 1979). If bed shear stress and turbulence are great enough material up to coarse-grained size in the aggraded bed may be fluidised by intergranular collisons and a moving rheologic layer of bed material

Figure 58: Aggradation in Cropp River following flood of 24 January 1980. Bed was raised by up to 1.5 m during the flood peak. During the falling stage and the following flood event all sign of terracettes left by aggradation was removed.

a

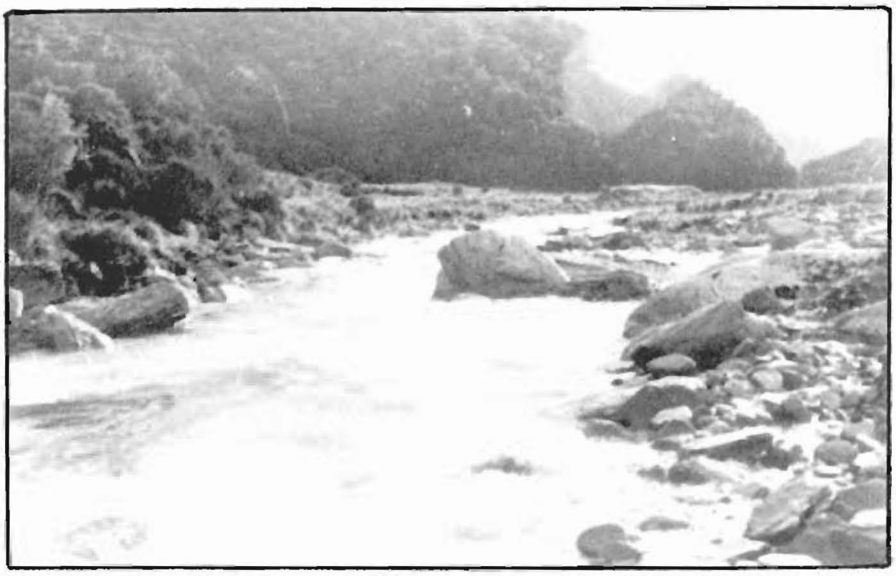
b

Figure 59: Aggregation during flood of 24 January 1980 formed a bed with an even gradient over considerable distances. Figure shows:

(a) reach under normal conditions; and

(b) reach with terracette (arrowed) showing how the pool and riffle structure of the stream channel was buried during an episode of widespread bed movement.

a

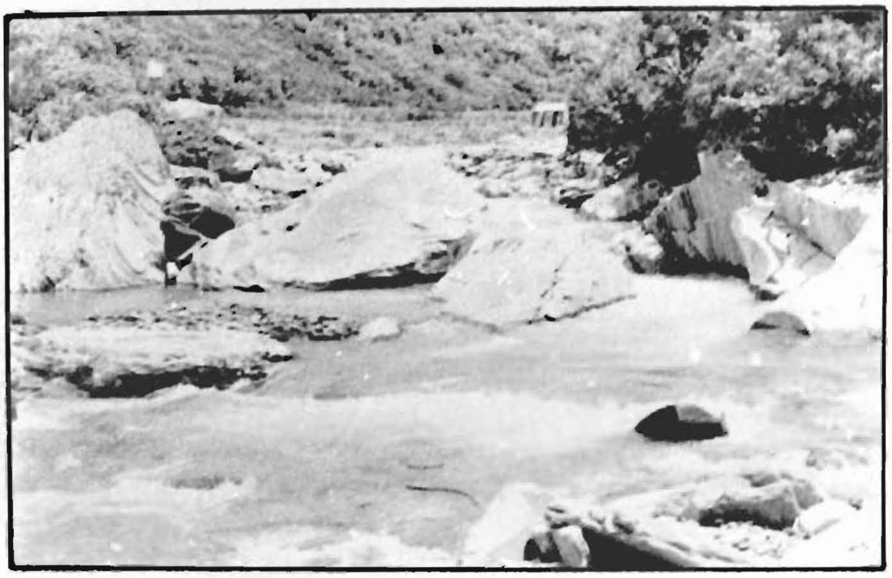


b

Figure 60: Large boulders were shifted during flood events of January 1980. Figure (a) shows the gorge in the Cropp below the gauging site on 26 January 1980. Figure b) shows the same view on 28 January 1980. Note the large boulder in the foreground of (a) is missing.

a

b



(Moss 1972) established. If this type of bed movement is established large boulders could be supported by dispersive grain pressure with energy input from the overlying sediment-laden supercritical water flow.

The amount of downstream movement produced by such a mechanism can only be speculated on as with the role of intergranular collisions in this debris flow-like movement, in grinding the rock material . However, such a mechanism might explain the downstream dispersion of boulders of suitable size to produce bed armouring material (Moss 1972), and it is possible that sufficient grinding, impact and abrasion occurs in a "rheologic layer" involving coarse poorly sorted material to produce the changes in bed material composition noted previously.

The small terracettes reflecting the maximum bed level during the storm of 24 January 1980 were widespread in the Cropp but were only preserved for a matter of days. Similar terracettes have been observed in many small alpine streams on both sides of the Main Divide implying that similar processes may be widespread at very high flows.

10.4 EROSION MONITORING

In order to observe primary movement of sediment on hillslopes and to give quantitative data with which to compare sediment transport rates and to observe the types of slope failure prominent in a high rainfall regime slope conditions in a small catchment were monitored for a year.

A small catchment (the Limpopo River, Fig. 61) was selected and areas where erosion appeared to be occurring were identified. These were of several types:

1 exposed non-vegetated ridges;
2 bedrock faces;
3 talus (scree-like) deposits;
4 streambank erosion;
5 incipient shallow regolith failures.

In each of seven sites, pegs were buried to depths of 50 cm and where some creep mechanism was anticipated, plastic hoses were pushed into the colluvium (using a steel bar to provide rigidity during emplacement). The bases of the erosion pegs were painted as was the surrounding ground as an indication of the ground surface position at the time of emplacement. Where rockfall movement was anticipated large squares were painted on bedrock surfaces and lines painted on screes to trace particle movements.

Figure 61: Limpopo River looking down into Danger Gully.

Observations were made at frequent intervals from when the erosion observation network was established (December 1979) until snow cover was complete (by May 1980). Further inspections were made in July, September, November and December 1980 by which time all sites were substantially snow-free.

Results of this study were inconclusive in primarily determining denudation rates for a small catchment. This was mainly because of the dynamic nature of the Limpopo catchment due to the high rainfall and the frequent high intensity rainfalls of the area. Despite the lack of consistent quantitative denudation data many interesting observations were made in studying this catchment.

Painted lines across talus slopes on erosion sites 1 and 5 (Fig. 62) were 60% removed within one week of installation. Painted material from site 1 was recovered from the debris dam in the Limpopo River (200 m from the source). This was not material from the scree but painted material from the bedrock face above the scree.

Observations of the morphology of the scree-like deposit of site 1 showed that during large storm events (return period greater than one month) loose material was washed from the bedrock face and removed from the scree in small debris flows and shallow (less than one metre) rills. Rills and debris flow lobes were flattened out during smaller rainfall events and by wind action. Any wind or rain could cause intermittent dislocation of small blades of schist from the bedrock face onto the scree.

Erosion stakes on bedrock did not generally remain in place over winter especially where exposed to spring avalanche activity and those that did showed no significant erosion. It is possible that freeze-thaw activity moved both the stakes and the surrounding rock.

There was complete burial of the stake network on site 2 by a small (10 m^3) debris flow deposit. The stakes remained unmoved by this event.

Squares painted on bedrock faces were all 95% removed within the year's observation period. As the bedrock consists of highly shattered schist which breaks readily into .5 cm thick blades up to 10 cm long the removal of the painted surface indicates a 2 cm+ retreat of the steep bedrock faces at the Limpopo over one year.

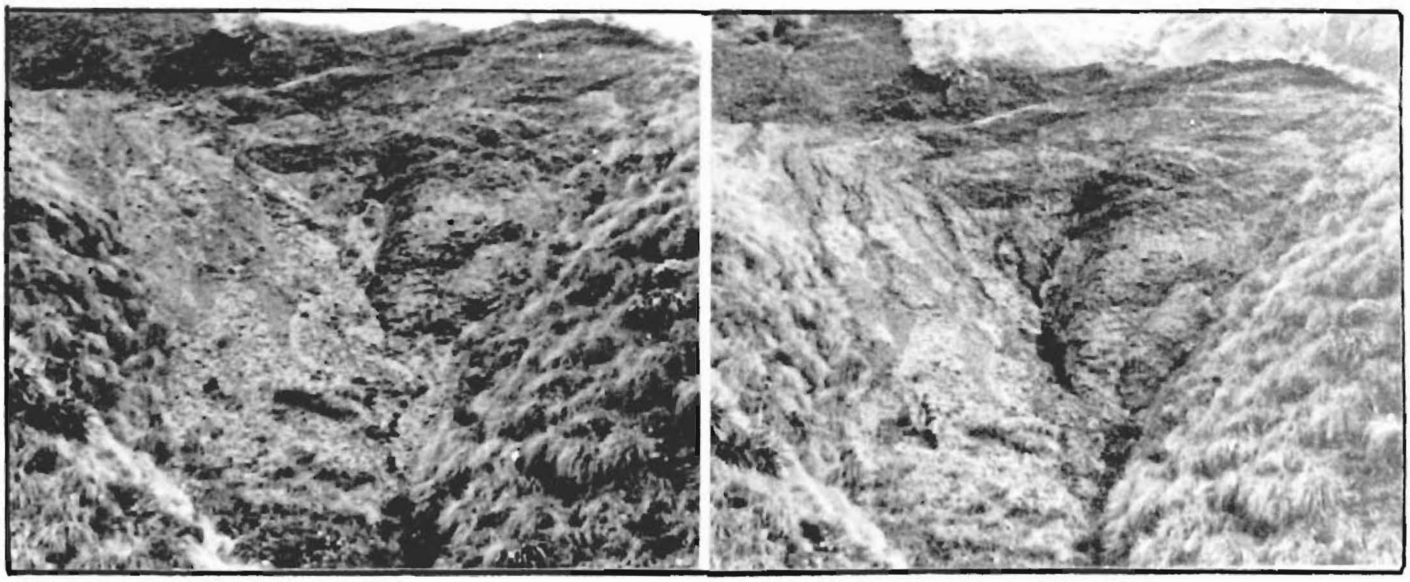
The site where soil creep was suspected (site 4) showed no measurable movement. However, observations in early summer indicated that subsoil drainage paths may influence soil creep, localise failure of the turf mat

Figure 62: Erosion monitoring sites, Limpopo River:

(a) Two shots of site 1 showing rilling of scree. The headwall of the site is undisturbed bedrock.

(b) Erosion site 5, this site moved by a series of small shallow rotational slumps, movement was in response to undercutting by the stream.

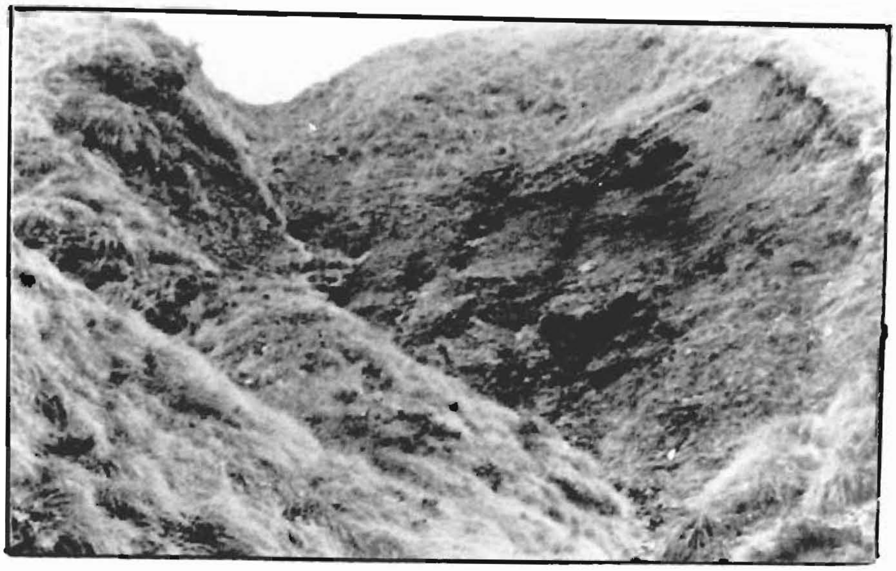
(c) Erosion site 7, exposed bedrock faces at the head of the Limpopo River. Areas of bedrock 0.5 m^2 were painted and within one month 60% of painted material had been removed. 98% removal occurred within eight months on all slope facets.



a

b

c



and, under the influence of avalanche conditions, sufficiently disrupt the soil to allow initiation of active erosion. These observations include disrupted turf mats with sub-surface "pipes" about 10 cm in diameter (Fig. 63) channelling flow back to the failure site. There was also a large volume of soil mat and vegetation material within and on avalanche debris (Fig. 64).

Plastic hoses buried in slope debris showed two types of creeping failure. At site 5 small rotational slumps one to two metres deep displaced the pipes without deforming them. The material in this deposit is poorly sorted and acts as a cohesive soil (probably because of included organic matter and soil derived clays). Much of the displacement occurred while the deposit was snow-covered.

At site 1 all four hoses showed distinct shear planes within the deposit parallel to the surface of the scree and to the imbrication of the schist blades. Two shear planes were identified, one at about 15 cm depth and the other at 40 to 50 cm (Fig. 65). Because these hoses were only excavated after the year's observation it is not known when the shearing occurred. A probable explanation is that the shallow shear represents normal gravity sliding of the surface material with lubrication during heavy rain and/or by frost heave and the deeper failure plane is a response to loading with (up to 5 m) snow cover during the winter.

Overall the study identified locally very high erosion rates but failed to quantify rates or to relate local erosion rates to catchment denudation. A study which would reliably quantify denudation rates in this situation would have to involve a number of small catchments and detailed systematic work to allow statistical analysis of results. Such a project would need to be undertaken over a long period of observation.

Figure 63: Soil pipes. Shallow failures of the soil mat are often associated with 10 cm$\pm$ diameter subsurface drainage channels (arrowed in b).

a

b

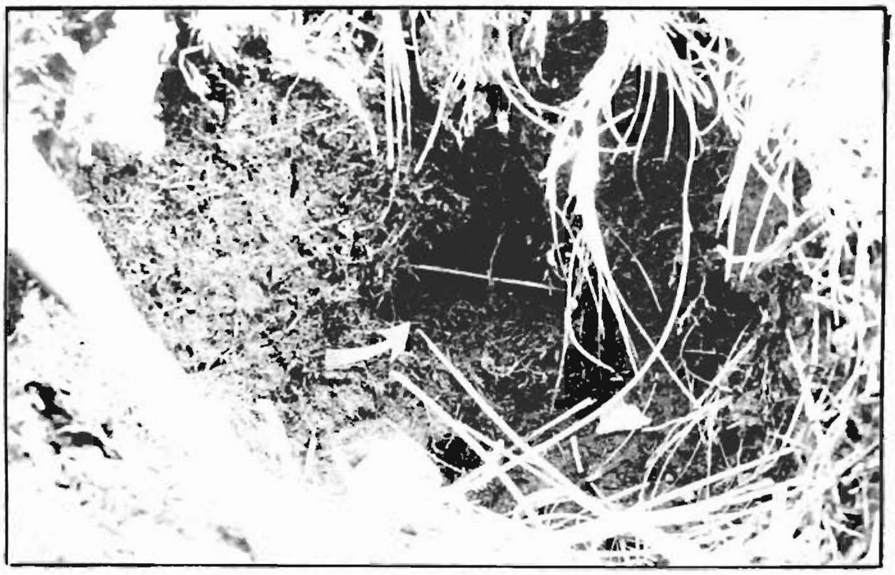


Figure 64: Whether or not winter snow cover inhibited sediment movement is unknown. However, spring avalanche activity removed considerable volumes of soil material from slopes in the Limpopo River.

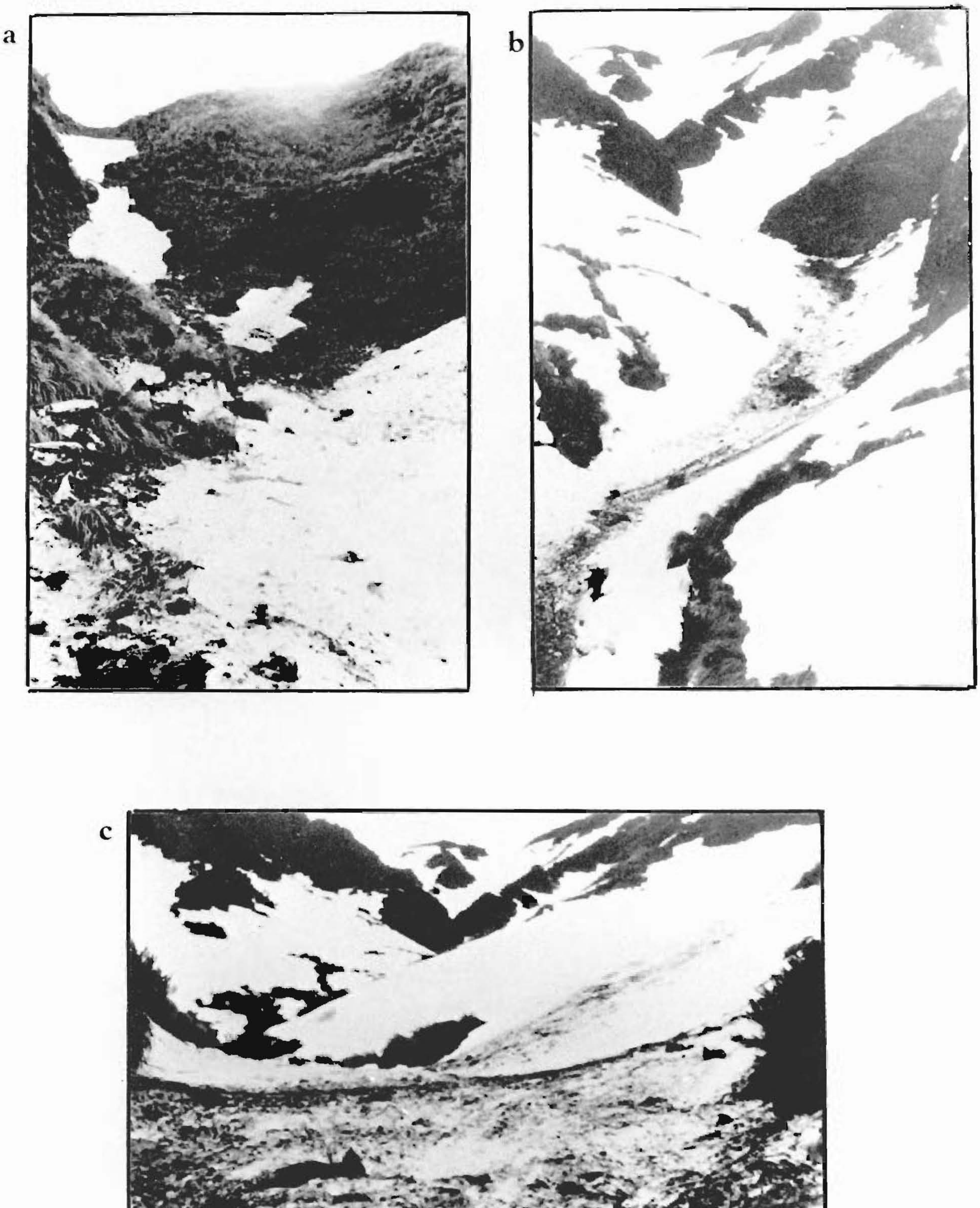
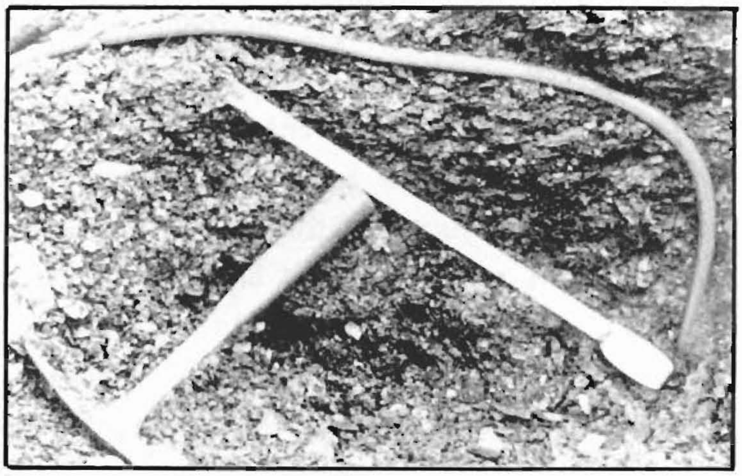
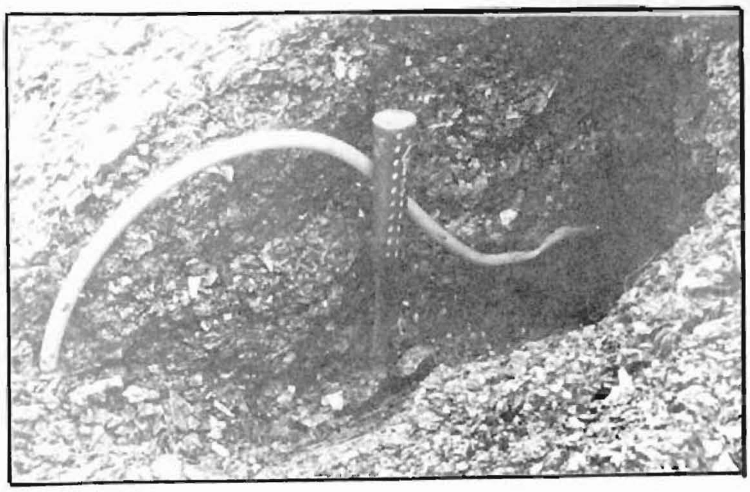
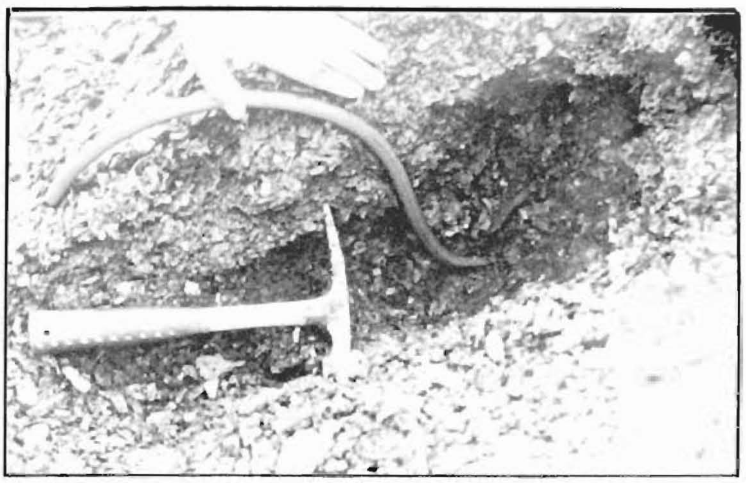


Figure 65: Buried hoses on erosion site 1 showing discrete planes of shear in scree-type deposit.


A

B



C

D


CHAPTER ELEVEN: GEOMORPHOLOGY AS AN INDICATOR OF EROSION HISTORY

11.1 GEOMORPHOLOGY

A late Pleistocene-Holocene glacial chronology for the Cropp area was established in Chapter Six. Two major glacial geomorphic events are recognised in the Cropp. Glacial planed bedrock surface forming ridge tops throughout the valley, are considered to be latest Pleistocene and to have been substantially formed by 12,000 BP.

Holocene ice advances between 8000 and 11,500 BP resulted in downcutting in the cirque basins and deepening of the main Cropp Valley. Following deglaciation (except for the high basins of Rotunda Creek and Beaumont Basin) a series of rapid changes occurred. Tarkus Knob is a bedrock promontory in the floor of the main Cropp Valley which was overridden by ice during the early Holocene glacial events and behind which a large hollow was scooped. Following deglaciation this hollow formed a lake which controlled the base level of streams and slopes upstream of it (and which was drained following fluvial downcutting through the middle gorge of the Cropp). In Rotunda Creek large talus cones developed on north and west facing slopes, graded to the lake-controlled base level. On the south side of the lower Cropp a series of large fans were formed at this time, probably by major rock falls from the oversteepened bedrock faces left after glacial retreat. Fan building lasted until the valley sides had graded to a V-shaped profile at which time fluvial downcutting became the dominant force of geomorphic change.

The present distribution of erosion features is related to post-glacial adjustments to a fluvial geomorphic system and to particular controls of climate and bedrock lithology and tectonic fracturing effects.

Areas predisposed to large-scale erosion are:

1 Glacially oversteepened rock faces where schistosity orientation makes the slopes unstable. These slopes fail either as gravity collapse structures or as rock fall (Figs 33 and 36);

2 Areas along the stream length where major undulations in the stream profile occur due to the uneven glacial floor profile. These are areas where the threshold of critical power of the stream (Bull 1979) is exceeded and valley degradation occurs (Fig. 66). Associated with degradation of the stream bed and backcutting of gorges and waterfalls, riverbanks are oversteepened and become an active sediment source.

a

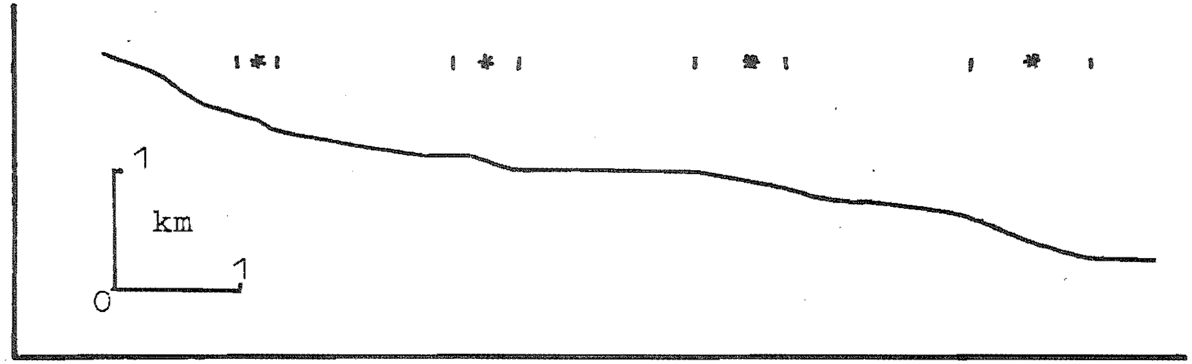


Fig.66:Long profile of the Cropp River.
* indicates gorges or waterfalls.From
N.Z.M.S.1,S64.

Figure 66 (ctd) The gorges of the Main Cropp River:

(b) Upper

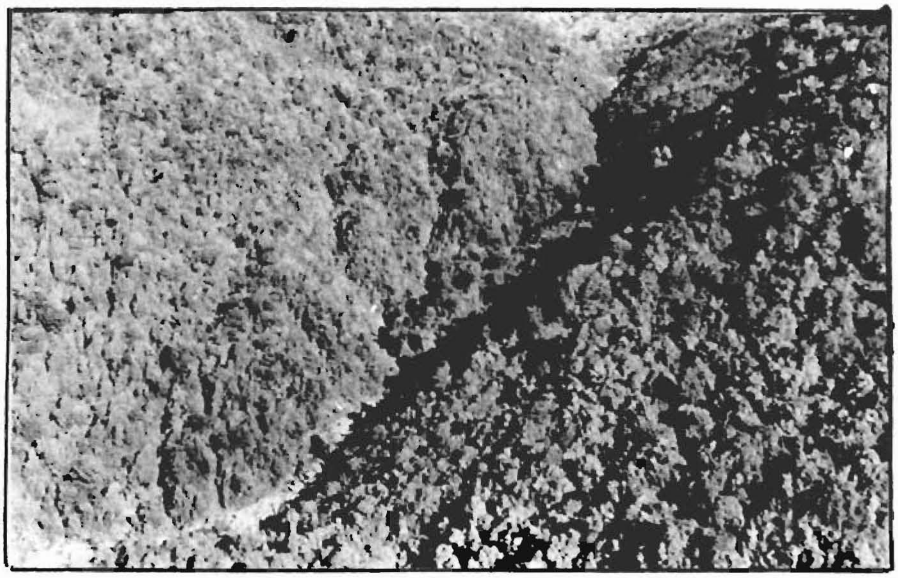
(c) Mid, and

(d) Lower gorges of the Cropp River.

b

c

d



3 In areas where bedrock is highly shattered (as with the Danger Gully areas, the central biotite zone of Chapter 3) the establishment of fluvial river profiles, both long profiles and V-shaped cross profiles, has been rapid. In these areas slope profiles grade directly into the streams and lowering of stream level can initiate erosion on hillslopes which is effective over the entire slope length forming a new graded profile. In these areas fluvial downcutting controls the rate of erosion as all slopes are potentially responsive to changes in river bed level.

The high level glacially planed bedrock slopes and the slopes in cirque basins such as the Top Cropp, Noisey Creek and Beaumont Basin are all affected by slower erosion processes less effective at modifying geomorphic features. Freeze-thaw movement of loose material, soil creep, avalanche scour and minor rilling and gullying occur in these areas. None of these erosive processes is as effective in terms of sediment supply as the mass movement phenomena promoted by fluvial undercutting of oversteepened slopes.

The net result of erosion on these low gradient surfaces is denudation without significant modification of the form of the slope. Denudation rates on these surfaces are probably relatively low. From the maximum height of tors left by differential erosion of friable and resistant lithologies on the glacial planed surface of Galena ridge (2.5 m) a minimum denudation rate of 0.2 mm per year for the past 12,000 years is implied (Fig. 67).

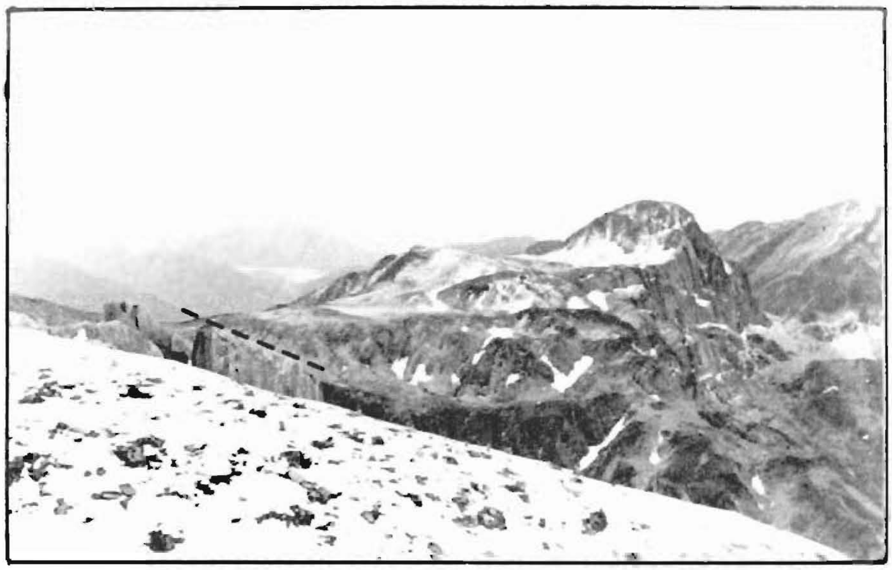
Geomorphic Regions

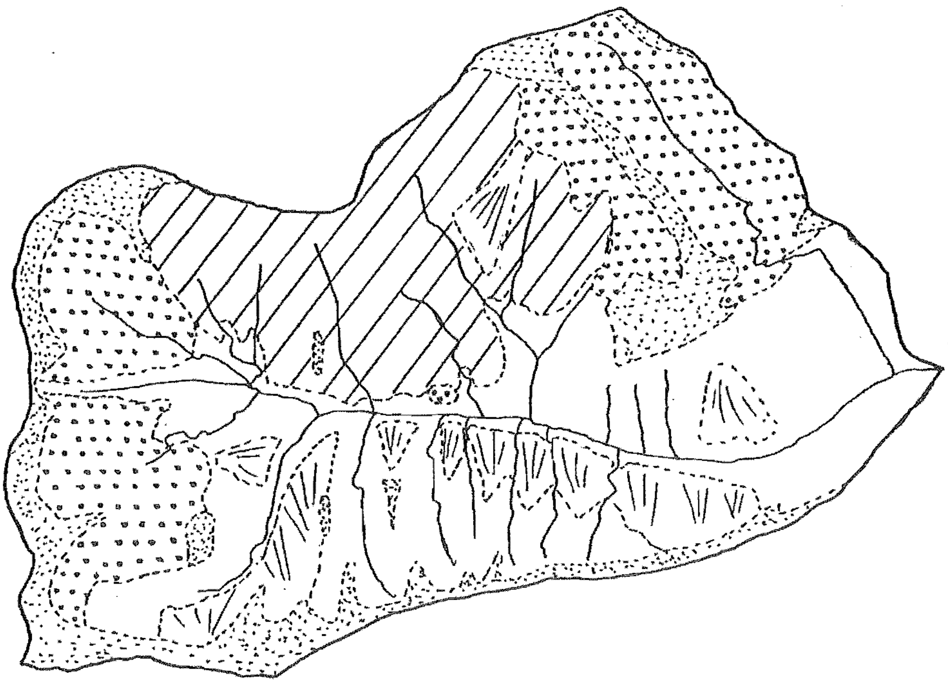
Five types of geomorphic region have been recognised in the Cropp catchment based on the type of slopes present and the amount and type of erosion indicated by the geomorphology (Fig. 68).

1 Flakey schist: these are areas where the bedrock is dominated by highly fissile, often shattered fine-grained pelitic schists. As described above for Danger Gully, in these areas hillslopes grade steeply and directly into the stream bottoms, most of which are actively cutting into bedrock. It is estimated that 95% of the ground area of these regions is susceptible to erosion resulting from a change in stream base level.

2 Fan: mainly on the south side of the Cropp Valley these regions are dominated by large fans formed following Holocene ice retreat by gravity collapse of coarse material from oversteepened slopes.

Figure 67: Tors on Galena Ridge preserve a relict surface and give a lower limit for denudation of the surrounding area since the time of formation of the surface.





Alpine
Cirque basins
Fan
Flakey schist
Valley cutting

Fig.68. Geomorphic regions; see text for explanation.

The fans represent considerable material in storage, much of which is subject to fluvial erosion at present.

3 Alpine: these are the regions where relict glacial surfaces are preserved on high ridges. Erosion was dominated by rock fall and gravity collapse following ice retreat but is now mainly freeze-thaw-wind and slope-wash dominated.

4 Cirque basins: these are relict Pleistocene ice-formed valleys deepened and widened during early Holocene glacial events. They comprise wide U-shaped valleys not yet influenced by headward erosion of the fluvial valleys. They have been preserved because of physical barriers to valley incision (the resistant ultramafic rocks which form a sill across the mouth of the Noisey Creek cirque (Fig. 31)), or insufficient runoff from the basin and too short a time for fluvial landforms and valley profiles to be established.

5 Main Valley cutting: this is the area of the Cropp where passage of water through rather than runoff from the area provides impetus for fluvial downcutting. Stream power far exceeds critical power (Bull 1979) in the long-term because of post-glacial adjustment of stream gradient. Erosion of stream banks and mass movement resulting from undercutting of slopes are the dominant erosional features in this region.

11.2 ESTIMATES OF HOLOCENE SEDIMENT YIELDS FROM POST-GLACIAL DOWNCUTTING

Hayward (1980) constructs a "paleo surface" in order to estimate the volume of material removed from the Torlesse Stream catchment over the last 2.5 Ma. Comparison of the calculated value with measured annual sediment yields is used to suggest that present-day erosion rates are lower than long-term rates.

In the Cropp it has been possible to create a paleo surface representing the level of glacial erosion at the end of the Pleistocene 12,000±1000 BP. This surface was constructed by joining relict glacial surfaces which are widespread through the catchment. Sixteen cross profiles of the present Cropp Valley, when compared with the "paleo surface" allow an estimation of the area, and volume, of material removed from 12,000 BP to the present for various parts of the Cropp catchment (Fig. 69 and Table 8).

Fig 69: Serial cross sections of the Cropp catchment showing present cross profiles and overlay showing profiles of the valley at 12000BP reconstructed by joining relict slope facets. Shaded area represents the material removed from the catchment by erosion and gives the basis for calculation of long term denudation rates.

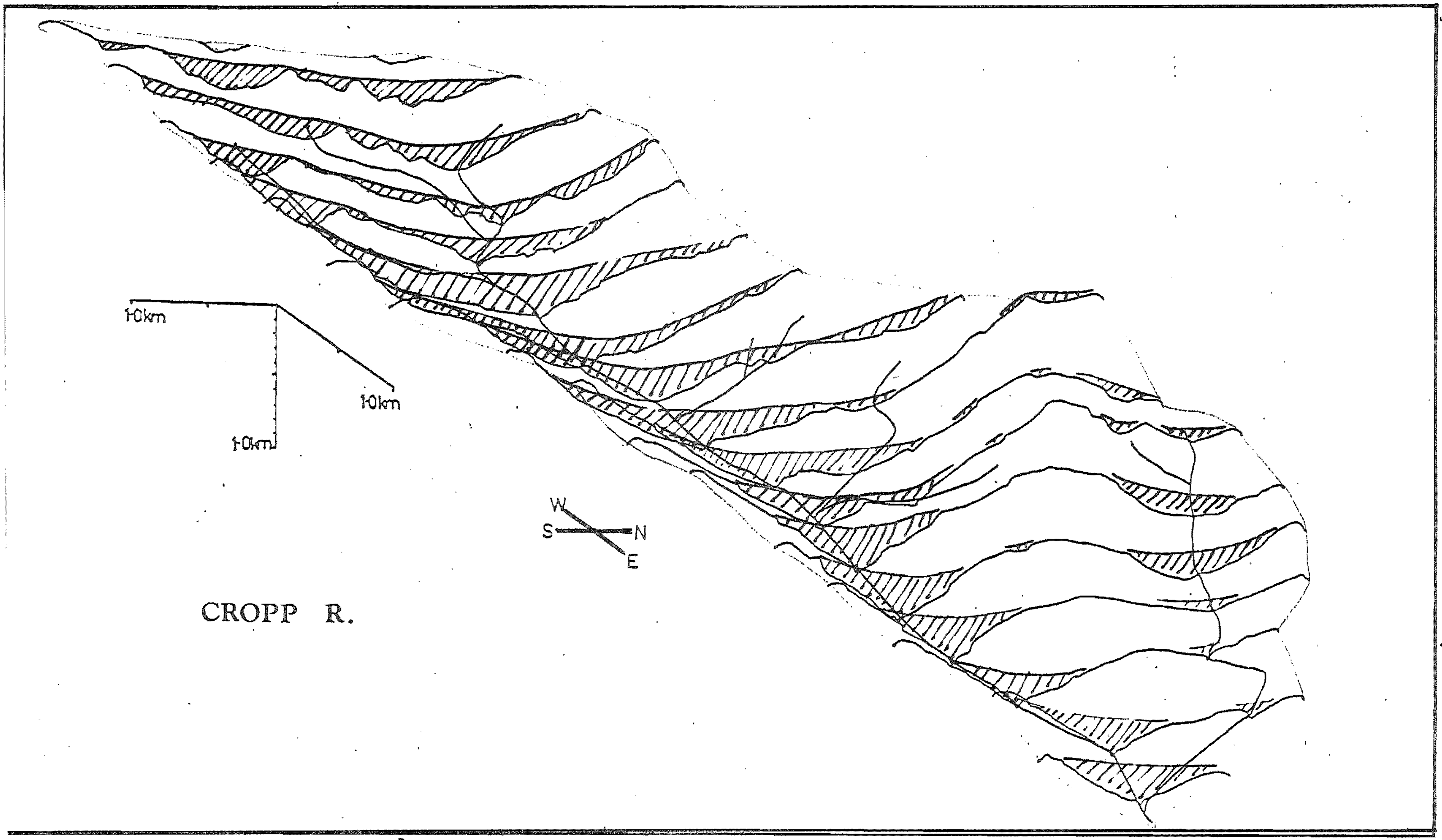

1·0km
1·0km
1·0km
W
S
N
E
CROPP R.

TABLE 8.

	TOTAL CROPP	TOP CROPP	NOISEY CREEK
Catchment area	$27km^2$	$12 \cdot 17km^2$	$3 \cdot 6km^2$
Volume of material removed (m^3)	$2 \cdot 7X10^9$	$1 \cdot 9X10^9$	$8 \cdot 21 \times 10^7$
Total denudation	103m	159m	22·8 m
Denudation rate 12000years	8·6mm/y	13·3mm/y	1·9mm/y

TABLE 8b

	%of area of Cropp	%of sed produced	DENUDATION RATE* for total Cropp	for upper Cropp
Alpine	20	1	0·45mm/y	0.2mm/y
Fan	10	9·7	8·66	6·65
Cirque	22	13·55	5·55	12·9
Valley cutting	30	60	18	27·34
Flakey schists	18	15·8	7·89	9·5

*Assuming total denudation rates of 9mm/yr and 14mm/yr

The overall denudation rate for the Cropp catchment is estimated at 8.3±3 mm per year for the past 12,000 years. Similar calculations for the upper Cropp catchment above the gauging site on the middle gorge gives a denudation rate of 13.3±3.4 mm per year (errors are estimated from the reliability of geomorphic data). This represents removal of 35,243±9,010 tonnes/ km^2 per year (assuming an initial rock density of 2.65 tonnes/m^3). This is very close to the measured 1980 suspended sediment yield of 32,410±11,290 tonnes/km^2 per year.

Examination of denudation data for the five geomorphic regions discussed previously (Table 8B) shows that considerable variation exists in the long-term denudation and erosion rates for the different geomorphic types. Denudation in alpine regions is insignificant; in cirque regions it is low overall although the recently active Beaumont Basin increases the value considerably, if only the upper Cropp is considered. Fan and flakey schist regions are very similar although it is likely that fan regions saw a burst of activity in the early post-glacial and are presently not as active as flakey schist areas.. Most importantly however, is the observation that main valley cutting has resulted in denudation at about twice the regional average.

The observation that valley deepening and widening is the most important agent of geomorphic change and the most important process producing sediment from the Cropp River supports observations made in Noisey Creek where these processes have been halted because of the presence of a belt of resistant ultramafic rocks across the valley mouth.

In Noisey Creek there has been no significant stream incision, in the Holocene, despite annual precipitation of 8500± mm per year. The denudation rate calculated from cross sections is 1.9 mm per year for the past 12,000 years. This represents a sediment yield of 5,035 tonnes/ km^2 per year compared with predicted yield (from McSaveney et al., in press), for a rainfall of 8300 mm, of 17,000± tonnes/km^2 per year.

In conclusion it seems that stream downcutting is the major control on slope processes and on sediment movement from sub-catchments in the Cropp area. The rate of stream downcutting is affected by runoff, sediment supply and geology. Resistant lithologic units provide a barrier to downcutting and suspend large-scale erosion in the upper catchment to a rate comparable with the rate of downcutting.

It is interesting to speculate what the sediment yield of a catchment similar to the Cropp in terms of geology and climate, but with a channel graded to a stable base level would be. The Noisey Creek value of 5,035 $\pm$ tonnes/km^2 per year is a possible indication.

CHAPTER TWELVE: IMPLICATIONS OF CROPP RIVER SEDIMENT YIELD DATA

The long-term denudation rate of the upper Cropp River is very close to the measured annual rate for 1980. The errors on both values are large but the values suggest that suspended sediment contributes the vast majority of the sediment load and that bedload and dissolved load contribute less than 20% of the total.

Both long-term and short-term denudation rates are very high (13.3$\pm$3.4 mm per year) with a long-term specific sediment yield of 35,245$\pm$9,010 tonnes/km^2 per year and a short-term rate of 32,410$\pm$11,290 tonnes/km per year for 1980. Combination of these values gives a yield of 32,685$\pm$11,565 tonnes/km^2 per year compared with a possible world record value from Taiwan of 31,700 tonnes/m^2 per year (quoted by Griffiths 1979).

Suspended sediment yield is a valid measure of erosion in the upper Cropp as suspended sediment comprises 80% to 90% of total denudation. Downstream effects of the very high sediment yield of the Cropp are minimal, as almost all sediment is moved in suspension, very little of which contributes to aggradation problems in the Hokitika River system. (This is supported by observations of gravel composition in the lower Hokitika which is nearly all greywacke gravel, with schist underrepresented considering the water yield of schist catchments, estimated at 35% of Hokitika water yield.)

Very high sediment yield values do not imply any danger to the downstream Hokitika River system as almost all sediment is carried in suspension and is normally passed straight out to sea within a flood event. However, as can be seen by the siltation which occurs behind the Westland Catchment Board hook groyne at Kowhiterangi any obstruction to normal flood discharge of the Hokitika River causes significant siltation. The volume of suspended sediment regularly produced by the Cropp and presumably also the Price River and Rapid Creek would produce major siltation problems for any structure which reduced gradients of the Hokitika River. The catchments of the Tuke and Dickson Rivers draining into the Mikonui could expect similar problems. One possible implication of this is that the dredge pond of the proposed gold dredge on the lower Mikonui River might encounter potential siltation problems during flood events.

The control geomorphic history appears to have on sediment yield and particularly the proportion of a catchment where streams downcutting to establish fluvial profiles following Pleistocene and Holocene glaciation may be significant in terms of producing predictive equations relating sediment yield to runoff, and in interpreting sediment yield data from individual catchments.

From a six-year study of sediment yield from the Torlesse Stream catchment (in the eastern Alps) Hayward (1980) calculated a specific sediment yield of 30 tonnes/km^2 per year for an area of 385 ha, with a runoff of 800$\pm$ mm/yr^{-1}. McSaveney *et al.* (in press) predict a specific sediment yield of 700 tonnes/ km^2 per year (for a runoff of 800 mm per year). This is consistent with an area of 16.5 ha contributing sediment at the predicted rate. If the Thompson and Adams (1979) estimate that schist catchments contribute sediment at three times the rate of greywacke catchment is valid, then the data could represent a sediment contributing area of 49.5 ha.

The Torlesse Stream data and the Cropp data are consistent with the hypothesis that catchment specific sediment yield is related to the geomorphic stability of the catchment; that areas where fluvial downcutting is occurring in the long-term are the sites of major sediment production; and that geomorphically stable slope/stream configurations (ie established scree fields in the Torlesse catchment) have minimal value as sediment producers.

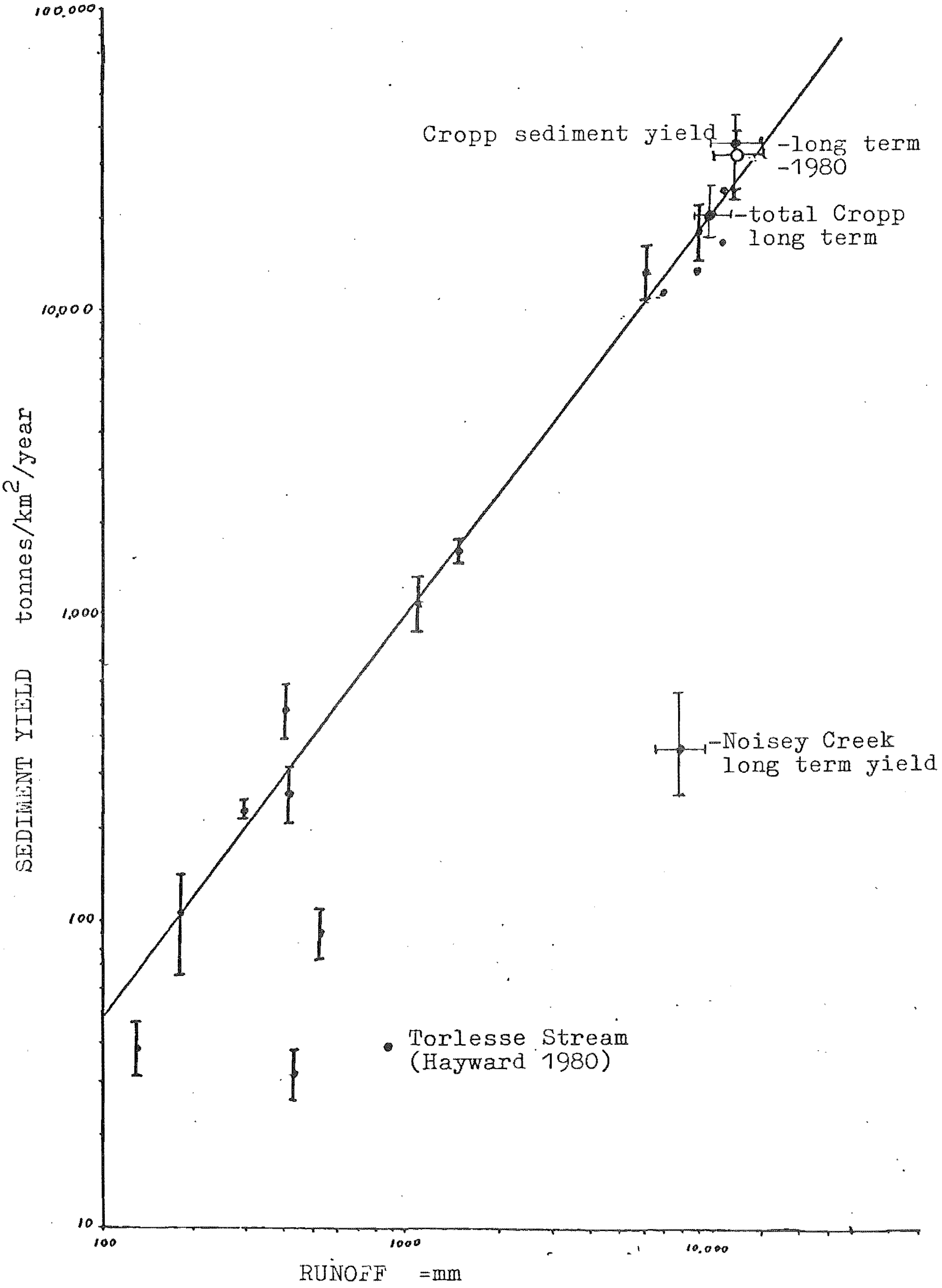
The above interpretation is generally supported by Hayward's (1980) work, (where 80% of the Torlesse catchment is covered by geomorphically stable scree) and by the high sediment yield at the Cropp which is affected by fluvial dissection on many scales and is a major yielder of sediment.

If other published sediment yield data is interpreted in terms of the proportion of the measured catchment which is in a state of geomorphic disequilibrium (ie where catchment geomorphology reflects drainage and erosion conditions relating to climatic or other control no longer operating, and is unstable under present-day climatic and drainage conditions) then predictive equations relating sediment discharge to runoff might become more meaningful.

McSaveney (1979), Griffiths (1979) and McSaveney *et al.* (in press) figure relations between specific sediment yield and catchment runoff. The most recent of these, McSaveney *et al.* (in press) (Fig. 70) slightly under-predicts the sediment yield for the upper Cropp River (although

Fig 70:Relation between sediment yield and catchment runoff,from M^{c}Saveney et al .(in press) including points for upper Cropp long term sediment yield,total Cropp and Noisey creek long term yields and Torlesse Stream sediment yield(from Hayward 1980).

FIG 70



100,000
10,000
1,000
100
10
SEDIMENT YIELD tonnes/km²/year
100
1000
10,000
RUNOFF =mm
Cropp sediment yield
-long term
-1980
-total Cropp
long term
-Noisey Creek
long term yield
Torlesse Stream
(Hayward 1980)

relation is within the measurement error limits) and predicts the yield of the total Cropp with considerable accuracy. Noisey Creek, however, is very anomalous as are plotted points for the Pomohaku, Manuherikia and Frazer Rivers.

I suggest that what is being considered in these equations is not a simple relation between runoff and sediment yield, but a complex relation involving geomorphic history. The variability in the plotted points can be explained if the degree of geomorphic disequilibrium is related to the time since deglaciation which is a function of geographic location, elevation, geology and precipitation.

While the McSaveney et al. (in press) equation may appear to predict sediment yields quite well, it does so not because the relationship between specific annual sediment yield and runoff is a fundamental one, but because the catchments samples represent a limited range of geomorphological conditions, in situations where effects of glaciation are related to precipitation. In fact the relation between sediment yield and runoff should be represented by two lines, an upper set of values where sediment yield is maximised because of geomorphological disequilibrium (the upper Cropp is probably a fair example), and a lower value representing the yield expected for a catchment where the base level is held constant and the catchment in geomorphic equilibrium (Noisey Creek and the upper Torlesse Stream, are probably close to this line). Depending on local glaciation history, tectonics, the presence of resistant lithologies (or vegetation effects) and man-induced geomorphic instability, many intermediate values between these two lines could be expected.

CHAPTER THIRTEEN: THE EFFECT OF BEDROCK GEOLOGY ON SEDIMENT YIELD IN AN ALPINE AREA SUBJECT TO EXTREME RAINFALL

The geology of the Cropp area reflects a geological history involving three metamorphic and associated structural events. Of the three metamorphic and deformational episodes the most recent is, at most late Cenozoic, and is probably continuing at depth at the present. Reflections of continuing deformation in the area are the prominent northeast and north-trending fault sets which control drainage directions over much of the area. A range of lithologies is present in the area with a range of responses and susceptibilities to erosive processes (in part determined by structural history).

It is the distribution of structural weaknesses and resistant lithologies which has determined drainage directions, patterns and stream morphology in the Cropp River.

The Southern Alps have been recognised as an area of active tectonism since the earliest geological studies. A review of geological and geomorphological evidence from within and outside the study area suggests that uplift has been occurring along the line of the Southern Alps since about 5 m.y ago and that the rate of uplift may be increasing. Uplift in the Cropp area is localised to the west on the Alpine Fault and to the east on a series of northeast-trending faults. The uplift rate in the Cropp region is close to the maximum in the Southern Alps occurring at 12.0±2 mm per year.

It is this high uplift rate which has prevented the Cropp area from attaining a state of geomorphic stability despite high erosion rates. Without the high uplift rate relief would have become subdued and climate would have ameliorated as a result, over a period of tens of thousands of years.

Independent estimates of sediment yield of the upper Cropp River result in similar very high values for a one-year period (32,400 tonnes/km^2 per year) and for a 12,000 year period (35,200 tonnes/km^2 per year). Within the limits of error on these measurements the short-term denudation is the same as the average for 12,000± years and is equivalent to a denudation rate of 13.2±3.4 mm per year.

In the Cropp River the major control on long-term sediment yield is the rate of fluvial downcutting. The more rapidly rivers incise their channels, the greater the instability of the stream banks and of slopes grading to the level of the stream channel, and the more hillslopes become dominated by fluvial landforms. Similarly the less the stream incision, the more relict surfaces will be preserved.

Geology is one of the main controls on the rate of fluvial downcutting, the presence of resistant beds slows the stream incision process and to a certain extent stabilises the catchment above the bedrock barrier. Conversely fault zones and weak lithologies promote downcutting and hence sediment yield.

The other control on the rate of fluvial downcutting along a channel is stream power. As defined by Bull (1979) stream power is the power available to transport sediment load defined by the stream discharge and channel slope. To this definition must be added a factor to describe the nature and amount of sediment load, as this affects the stream's capacity to erode and transport further sediment. These factors are functions of catchment runoff (rainfall), geometry and geology (in terms of the type of sediment transported by the stream).

Of these rainfall provides the energy input to the system providing both the medium and the mechanism (runoff) for sediment transport. Geology and catchment geometry are the controls which determine how, where, and if, the energy input from rainfall will be expended in erosive processes.

The geometry of the Cropp catchment is largely a relict of an early Holocene glacial geomorphology with few, broad cirque basins feeding a valley with a highly stepped profile. Some parts of the catchment which are near the main stream and not geologically unstable have, as yet, resisted fluvial dissection and valley entrenchment because insufficient time has been available since glaciation for deepening of gorges and backcutting of waterfalls. These catchments and those isolated by the presence of resistant rock units have preserved a high proportion of relict surfaces and show relatively few signs of fluvial dissection as the base levels of the upper catchments are stable and the impetus for slope modification is lacking.

Sediment yield patterns over the Southern Alps broadly relate to rainfall because of the energy input required to transport sediment and it is this effect which is measured by relations such as those of McSaveney et al. (in press), McSaveney (1979) and J Adams (1978). This simple relation breaks down where the transport mechanism does not rely on fluvial energy (as in catchments where debris flow is the major mechanism for sediment transport (ie Pearson (1980) gives figures which suggest a denudation rate of 24,000± tonnes/km^2 per year for a catchment with an annual runoff of about 1000 mm where sediment transport is by debris flow mechanism) or where sediment supply is limited by basin, channel and hillslope geometry or bedrock type (for example Noisey Creek or Torlesse Stream).

This relation assumes that the majority of sediment is derived from channel deepening processes and not soil erosion on hillslopes. This is in line with findings in Chapter Twelve, in Hayward (1980), and Tonkin et al. (1981). The contribution of soil erosion to sediment yield may be relatively small in a catchment like the Cropp, except where it is associated with mass movement features. This is supported by observations in the Cropp of overland flow on steep slopes during a high intensity rainfall event (40 mm± per hour) which carried no observable sediment load.

Other controls on sediment yield normally considered in studies of catchment stability, such as effects of vegetative cover and man- or grazing animal-induced instability, are probably not significant in catchments like the Cropp. The cause of the high sediment yield of the Cropp River is the combination of a stream with an uneven, long profile which is susceptible to downcutting with a high rainfall which provides the energy for removal of sediment from areas adjacent to stream channels, despite a healthy, essentially ungrazed vegetative cover over all but the high Alpine portions of the catchment.

It is considered likely that most of the sediment moved out of the Cropp is eroded from bedrock faces, transported by gravity-induced, water-lubricated mass movement to a channel and initially transported in steep channels as bedload. The friable schist which comprises most of the bedrock of the Cropp then undergoes various processes of comminution with the result that most of the sediment is removed from the catchment as suspended load. Thus bedrock geology controls the type of sediment produced from the catchment. The susceptibility of pelitic schist to comminution explains the high suspended sediment load of the Cropp. In contrast Hayward's (1980) study of the Torlesse Stream catchment showed a dominance of bedload from a catchment dominated by greywacke, which is more resistant to abrasion than the schists of the Cropp (in a stream with a relatively low channel gradient and therefore a low energy expenditure during sediment transport).

The estimates of uplift and erosion from the Southern Alps of J Adams (1978) (quoted by Walcott (1979), Wellman (1979), Howard (1979), J Adams (1979) and J Adams (1980)) are not compatible with the findings of this study. Adams calculates the amount of material involved in tectonic uplift, river load and offshore deposition for the Southern Alps and concludes that the three processes are in equilibrium with denudation rates equalling uplift rates. While the conclusion is supported for the Cropp region by the results of this study the uplift and denudation rates calculated for the Cropp (12.2 ± 2 mm per year and 13.2 ± 3.4 mm per year) are not even within the error limits of Adam's estimates (uplift maximum of 22 mm per year and errors on the uplift estimate of $\pm3.2\%$).

Bedrock geology in an alpine area with extreme rainfall has three effects on sediment yield.

1 It controls the patterns of sediment supply as a function of rock strength and therefore erodibility;
2 It controls the sediment transport mechanism by determining the grain size of material available for sediment transport; and
3 Geological uplift ensures a long-term continuity of supply of erodible material to the high uplift, high rainfall, high denudation area of the western Southern Alps.

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APPENDIX I

K/AR DATING

During August 1980 samples from six sites scattered through the study area were dated by the K/Ar method at the Institute of Nuclear Sciences (DSIR), Lower Hutt. The dating was done under the supervision of C Adams and J Gabites using techniques outlined in Adams (1975) and as modified for Adams and Nathan (1978).

The basic technique used is as follows:

Bulk samples of fresh rock were collected, broken up with a hammer and crushed in a ring mill. The powder produced was then sieved and the 1.25 to 2.35 Ø (210-420 micron) fraction was washed, dried and some retained as a whole rock sample. Mineral separates were obtained from the remaining sample using a Franz isodynamic separator, into various magnetic grades. The slightly magnetic fraction (settings 20^o, 20^o and 0.35 amps) comprised varying proportions of biotite and chlorite (the biotite in which was further concentrated in DS3 and DS5) and the very slightly magnetic (20^o, 20^o at 1.2 amps) was largely feldspar. These samples were then taken for analysis.

Percent K determination was achieved by acid digestion of powdered samples using hydrofluoric and nitric acids. The precipitate formed was redissolved and transferred to an ion exchange column (containing Biorad AG 50 x 12 cation exchange resin) and eventually eluted with nitric acid. The K+ solutions produced were analysed on a modified Unicom SP900 flame spectrophotometer. K analyses were duplicated and agreement within 1% was required between duplicates or a further pair were analysed.

Argon^{40} analysis was achieved using an isotope dilution technique in a mass spectrometer. A small powdered sample was fused under high vacuum by induction heating to about 1500^oC. A known quantity of Ar^{38} was then "spiked" into the system. Condensable gases were removed from the system in a cold trap and chemically active gases were pumped clear of the system using a titanium sublimation pump. The system was cleared of all non-inert gases by rapid heating of a tungsten filament. The radiogenic Ar^{40} plus Ar^{38} were then introduced to an AEI MS10 mass spectrometer. Comparison of peak heights for the known volume of Ar^{38} spike and Ar^{40} produced from a known sample weight allows the amount and then concentration of Ar^{40} in the sample to be calculated.

Both K and Ar^{40} analyses were calibrated with the international standards Arvonia slate (A7) and Ga BSO biotite at various times to check both accuracy and precision.

Calculation of ages used decay constants for K^{40} of

$${}_{B} = 4.962 \times 10^{-10}\ yr^{-1}$$

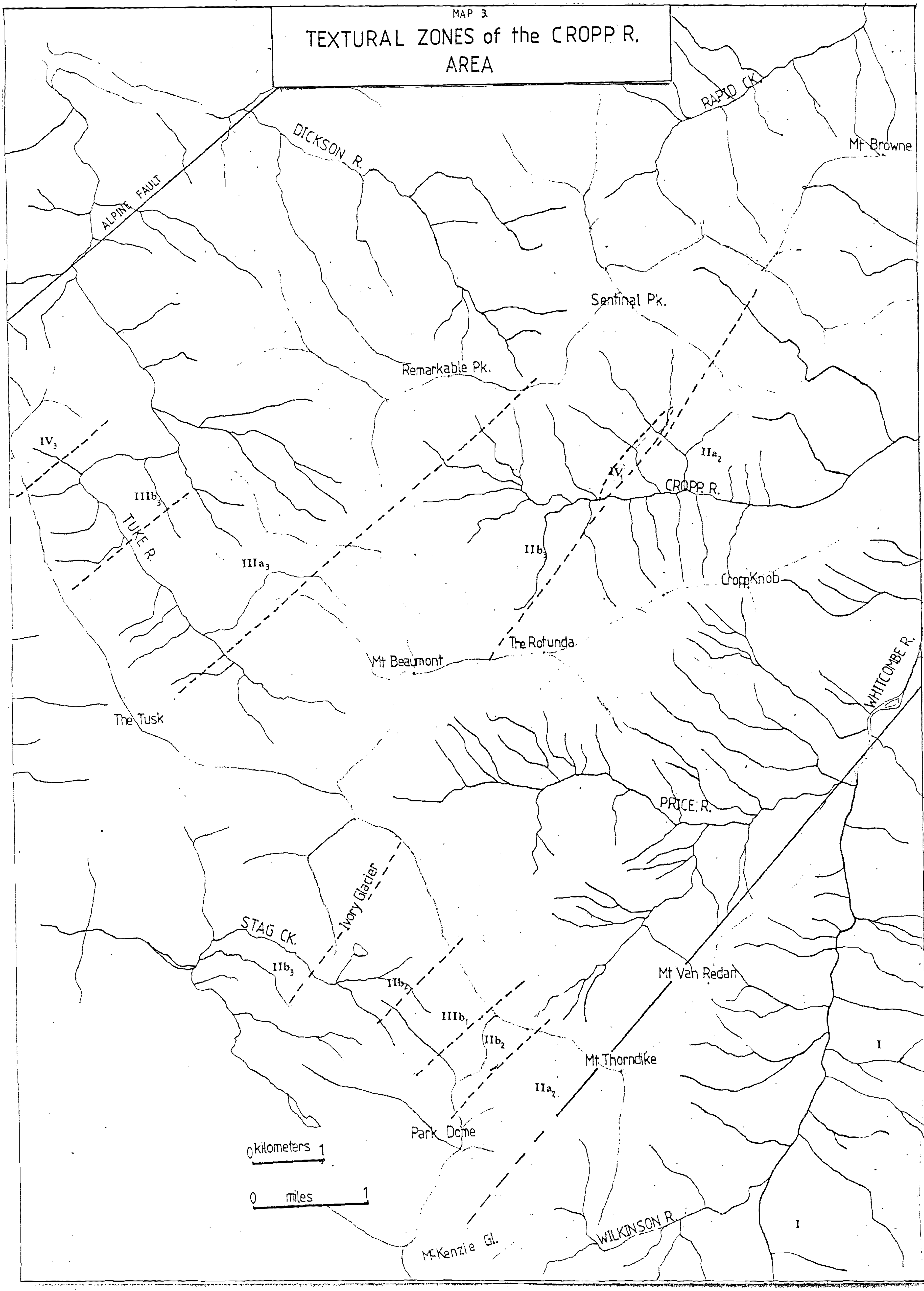
and

$${}_{e} = 0.581 \times 10^{-10}\ yr^{-1}$$

with an isotopic abundance ratio of

$$^{40}K/K = 0.01167 \text{ atomic } \%$$

Results are presented in Table 3 and discussed in Chapter 3, section 3.5.


MAP 3.
TEXTURAL ZONES of the CROPP R. AREA
ALPINE FAULT
DICKSON R.
RAPID CK.
Mt Browne
Sentinal Pk.
Remarkable Pk.
IV$_3$
IIIb$_3$
TUKE R.
IIIa$_3$
IIa$_2$
IV$_1$
CROPP R.
IIb$_3$
Cropp Knob
The Rotunda
Mt Beaumont
WHITCOMBE R.
The Tusk
PRICE R.
Ivory Glacier
STAG CK.
IIb$_3$
IIb$_2$
IIIb$_1$
IIb$_2$
Mt Van Redan
Mt Thorndike
IIa$_2$
I
Park Dome
0 kilometers 1
0 miles 1
WILKINSON R.
McKenzie Gl.
I